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Carbonate diagenesis and depositional cycles of the Mission Canyon limestone Madison Group of southwestern Montana

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CARBONATE DIAGENESIS AND DEPOSITIONAL CYCLES OF THE MISSION CANYON LIMESTONE, MADISON GROUP OF SOUTHWESTERN MONTANA

by

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ABSTRACT

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Carbonate diagenesis and depositional cycles of the Mission Canyon Limestone, Madison Group of southwestern Montana (129 pp.)

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The Mission Canyon Formation in southwestern Montana is a thick-bedded to massive sequence of limestone and dolomite, with very minor amounts evaporites and terrigenous clastics. Data collected from four stratigraphic sections shows that the rocks accumulated in a shallow sea of normal salinity. Paleontologic and lithologic diversity is relatively low, with crinoidal packstones, oolitic and crinoidal grainstones, and dolomite being the most abundant rock types. Diagenetic modifications include silicification, cementation, compaction, and dolomitization. In addition, some of the rocks have been extensively recrystallized. Cement fabrics indicate that cementation occurred in the meteoric phreatic zone. Most of the dolomite is of a penecontemporaneous supratidal origin, and was formed utilizing a local source of carbonate. The process of dolomitization resulted in the formation of a considerable amount of porosity, mainly as intercrystalline pores between dolomite rhombs, and as molds and solution enlarged molds of crinoids.

The four sections measured delineate a carbonate platform with an offshore oolitic shoal. Several shallowing-upward cycles are present; each generally consisting of sediments indicative of an open marine environment shallowing to a shallow subtidal zone, a tidal zone, or a supratidal zone.

Two distinct type of breccias are present throughout the study area: karst breccias, and evaporite-solution breccias.

A karst topography developed at the top of the section during the middle to late Mississippian accounts for the extremely variable thickness of the Mission Canyon Formation.

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INTRODUCTION

General Stratigraphy

The Mississippian Madison Group in the northern Cordillera is a marine sequence of predominantly carbonate rocks with subordinate amounts of terrigenous clastics and evaporites. It consists of two formations: the Lodgepole Formation below, and the Mission Canyon Formation above. The Lodgepole Formation can be further subdivided into the Cottonwood Canyon Member, the Paine Shale Member, and the Woodhurst Limestone Member. It is not certain that the Mission Canyon Formation can be adequately subdivided in Montana; it has been suggested that it be divided into an unnamed lower limestone and an unnamed upper limestone, separated by a thick bed of intraformational breccia (Sando, 1972), or into an unnamed lower member and an upper member that may be equivalent to the Charles Formation found in the subsurface of central and eastern Montana (Roberts, 1966).

In southwestern Montana, the Lodgepole Formation rests unconformably on the Sappington Member of the Three Forks Formation, and is conformable with the Mission Canyon Formation. Either the Big Snowy Group (Mississippian) or the Amsden Formation (Pennsylvanian) rests unconformably on the Mission Canyon Formation. There is no evidence for a major unconformity within the Madison Group (Norton, 1956; Sando and Dutro, 1960).

The Madison Group varies considerably in thickness throughout Montana. It ranges from a maximum of approximately 640 m thick in the Williston Basin of eastern Montana to about 550 m in the Big Snowy Trough of central Montana, and again to about 600 m in the Miogeosyncline of western Montana (fig. 1 and 2). It then thins gradually to about 200 m in both north and south central Montana (Smith and Gilmour, 1979). Some of the thinning that takes place to the south is a result of depositional factors, while most of the thinning that takes place to the north is a result of pre-Jurassic erosion (Smith and Gilmour, 1979).

Purpose

Although the structure and stratigraphy of the Mission Canyon Formation in southwestern Montana is relatively well-documented, the formation has not been studied in detail, particularly with respect to microfacies analysis. As the Madison Group is an important aquifer and a significant reservoir of oil and gas in Montana and elsewhere, a detailed microfacies analysis is proposed to define the depositional environments, diagenetic history, and modes of origin and development of porosity of the Mission Canyon Formation in southwestern Montana.

The specific area of study is bound by Interstate 15 on the west, the Montana/Idaho border on the south, Interstate 90 on the north, and Yellowstone National Park on the east (fig. 3).

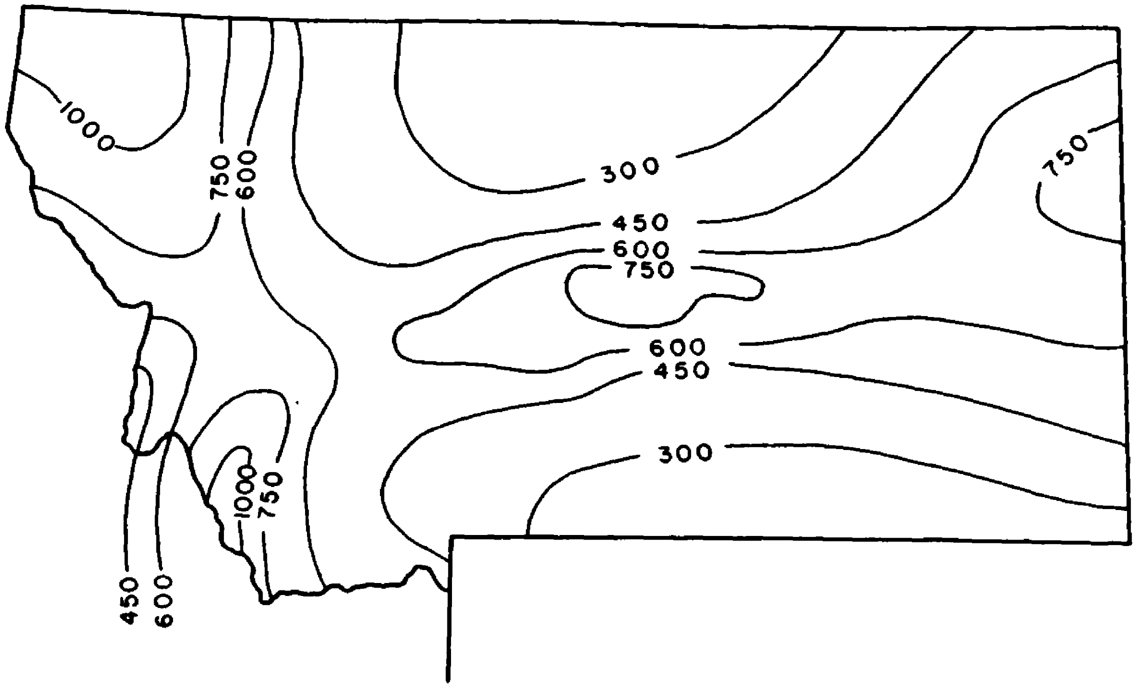


Figure 1. Isopach map of Madison Group in Montana. Thickness in meters. Modified from Sloss (1950), McMannis (1955), and Sando (1976).

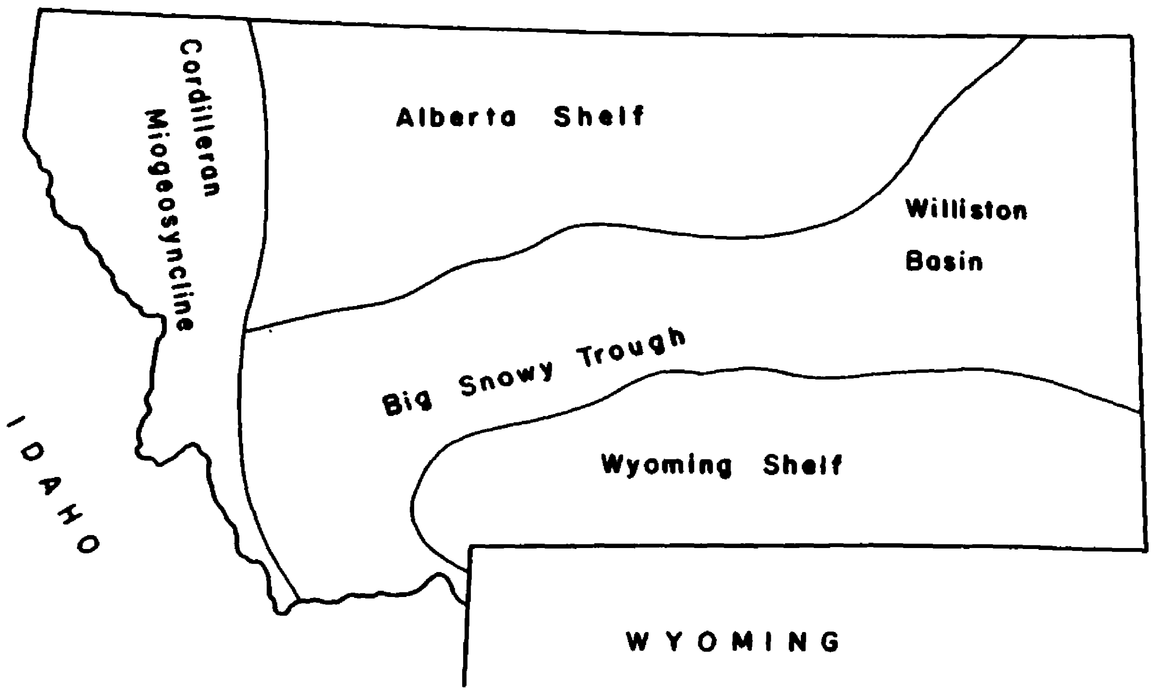


Figure 2. Major tectonic elements in Montana during the Early to Middle Mississippian. Modified from Sando (1976) and Smith and Gilmour (1979).

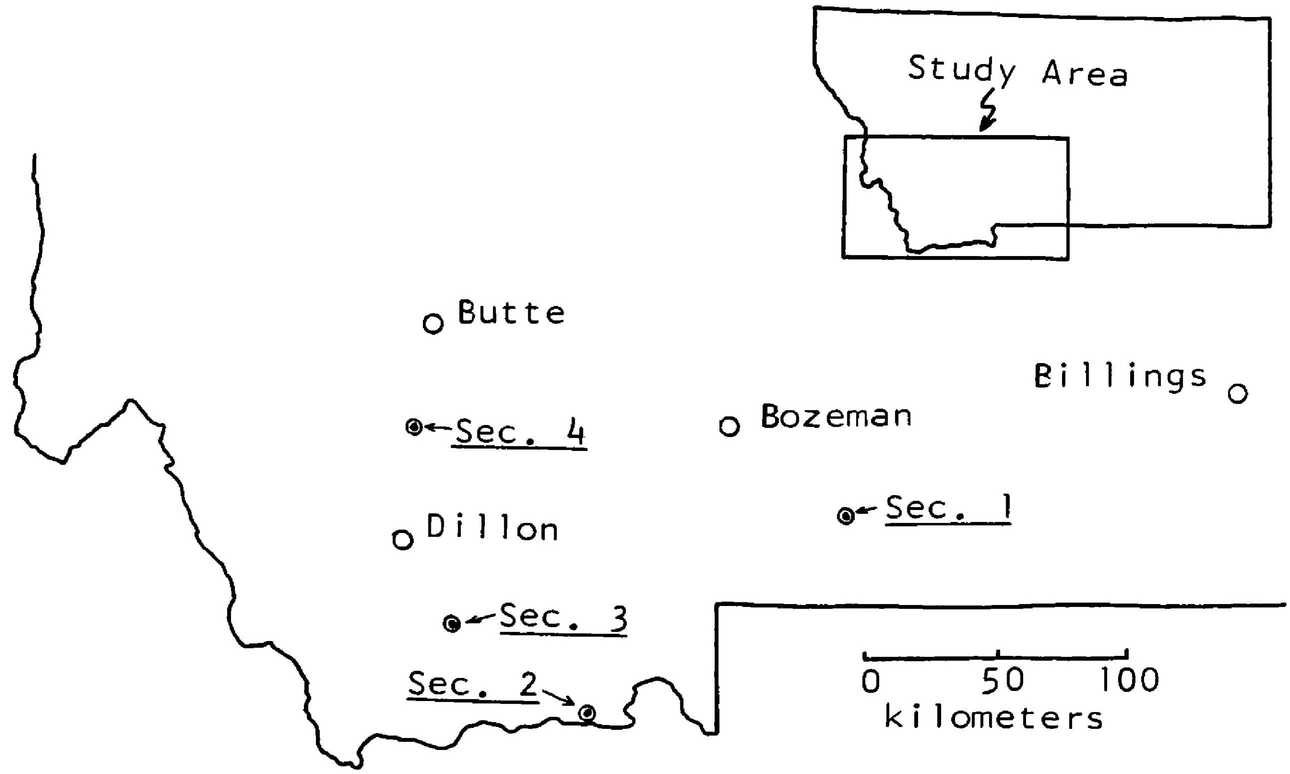


Figure 3. Map of study area showing location of measured sections.

Study Methods

This paper is based on the study of four stratigraphic sections of the Mission Canyon Formation examined during the summer of 1983. Sections were measured with a Jacob's staff and steel tape, and samples were collected at every break in lithology or every 3.3 m. The locations of the stratigraphic sections are given in the appendix and are shown in figure 3 above.

All rock samples were sawed, slabbed, and polished for examination under a binocular microscope. Approximately 100 representative samples were cut as thin sections, and acetate peels were made of an additional 75 samples following the methods of McCrone (1963). X-ray diffraction patterns of certain samples were analyzed to determine calcite, dolomite, quartz, and clay content (Diebold et al., 1963), and percent dolomite nonstoichometry (Lumsden and Chimahusky, 1980). Using the above techniques, rocks were examined for texture, composition, alteration, porosity, and diagenesis. Classification used is from Folk (1962) and Dunham (1962). Porosity nomenclature is from Murray (1960) and Choquette and Pray (1970). Diagenetic principles and terms are from many sources and are referenced in the text where appropriate.

Previous Work

Until recently the Madison Group, and in particular the Mission Canyon Formation, has been subject to relatively little intensive study in southwestern Montana. Hamblin (1939) and Sloss and Hamblin (1942) did a study of insoluble residues of the Madison Group and determined that this method was of practical value in differentiating the stratigraphic units. Sloss and Moritz (1951) summarized the Paleozoic stratigraphy of southwestern Montana, but they did not work specifically on the Mission Canyon Formation. Hall (1952) studied the Mississippian of northwestern Montana, as did Sloss and Laird (1945). Holland (1952) compared the Madison type section at Logan, Montana, with a section in northern Utah, although he only worked with the Lodgepole Formation. Stocker (1954) expanded Sloss and Hamblin's (1942) study into the subsurface of eastern Montana, and Andrichuk (1955) worked on the Madison of southcentral Montana and northern Wyoming. Sando and Dutro (1960) and Foster (1963) investigated the biostratigraphy of the Madison Group in Montana, and Middleton (1961) and Roberts (1966) worked on the karst and solution breccias in the Mission Canyon Formation. Most recently, Sando (1972, 1974) and Sando and Mamet (1974) examined the Madison Group in Montana and Wyoming, while Moore (1973) did a study of the Lodgepole Formation in southwestern Montana, and Sando (1967, 1976), Rose (1976), Smith and Gilmour (1979), and Peterson (1981) worked on the Mississippian in the northern Cordillera.

MADISON GROUP STRATIGRAPHY

Regional Tectonic Elements

There were several major tectonic elements active during the early Mississippian in Montana (fig. 2). These include the miogeosyncline, trending north-south through the extreme western part of the state, the Big Snowy Trough (unstable shelf) trending east-west through the central part of the state, the stable shelf to the north and south of the Big Snowy Trough (called the Alberta Shelf and Wyoming Shelf, respectively), and the Williston Basin in eastern Montana (Smith and Gilmour, 1979).

Considering these tectonic elements, Scholten et al. (1955) described, and Sando (1967) enlarged upon, the idea of referring to the rock units based on their depositional environment in the above setting. Their "Idaho Province" is synonymous with the miogeosyncline, the "Wyoming Province" is the very shallow shelf that is essentially the Wyoming Shelf, and the "Montana Province" is the somewhat deeper shelf that includes the Big Snowy Trough (Sando, 1967).

Depositional History

During the early Mississippian, the seas, which had withdrawn to the west in the late Devonian, began transgressing into southwestern and parts of south central Montana, depositing the black shales and siltstones of the Cottonwood Canyon Member of the Lodgepole Formation (fig. 4 and 5).

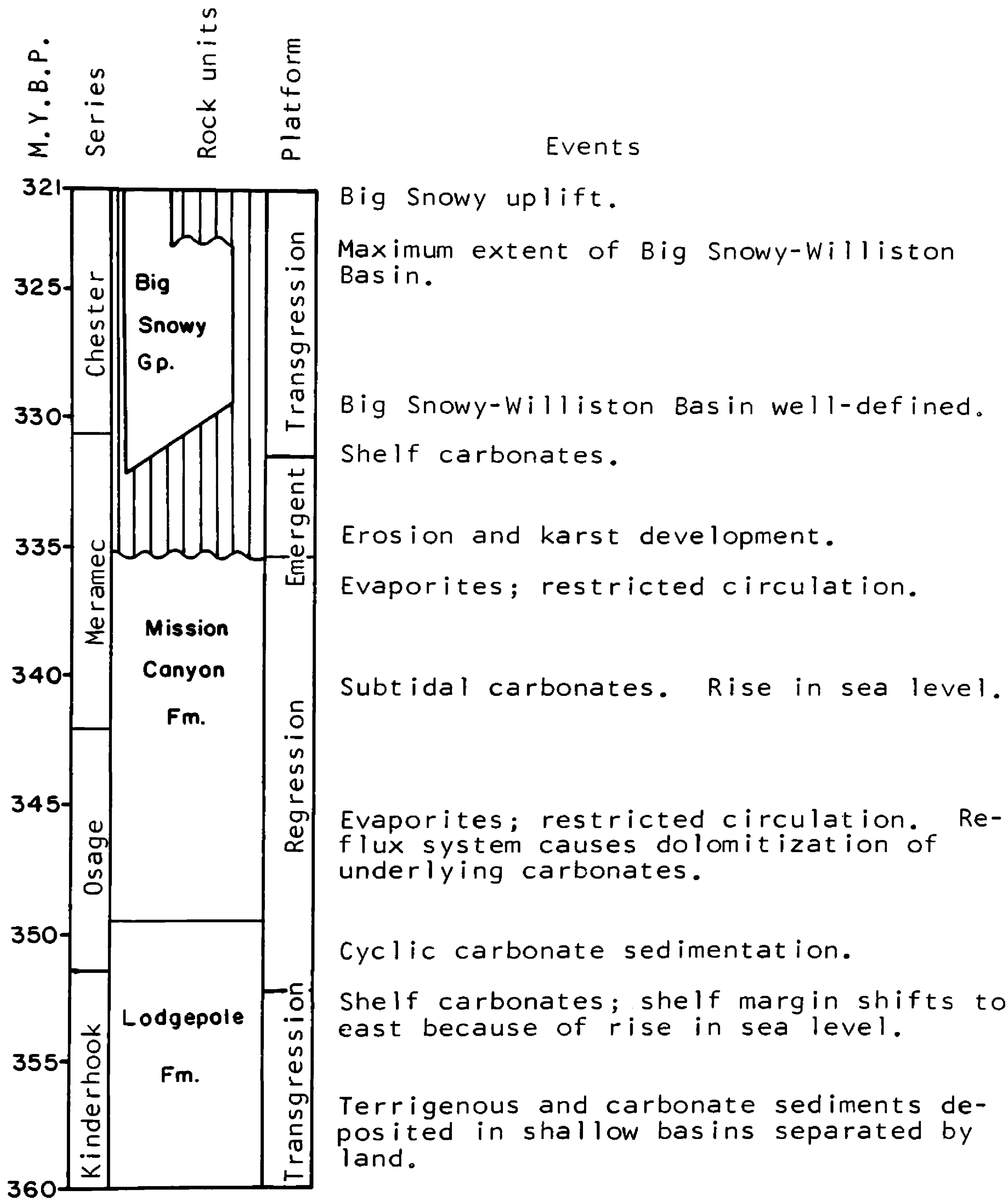


Figure 4. Stratigraphic column for the Mississippian in southwestern Montana. Modified from Sando (1976).

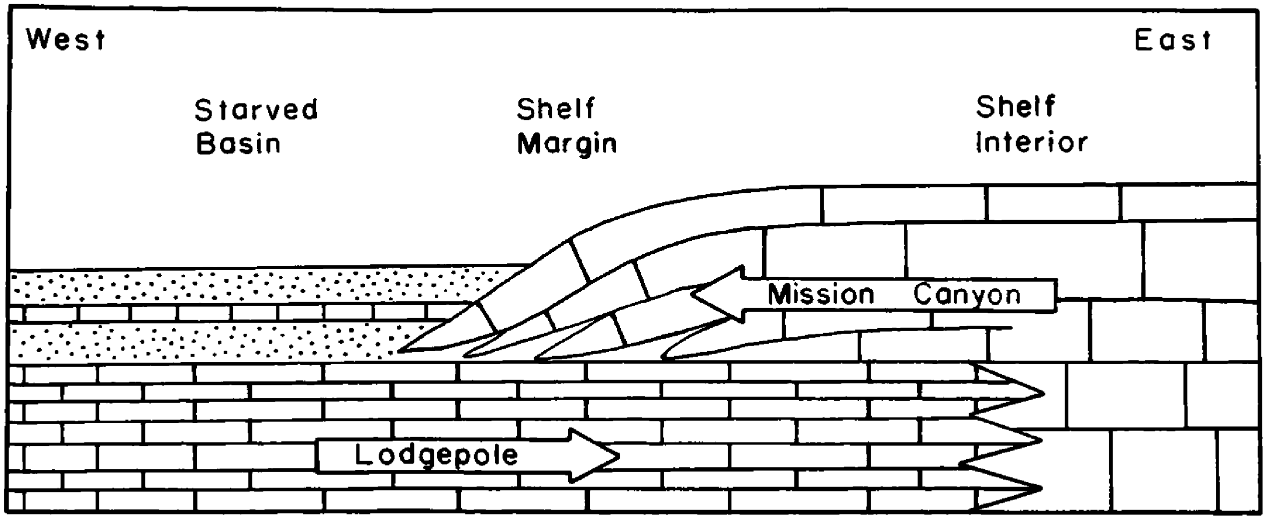


Figure 5. Stratigraphic model across Idaho and Montana, with arrows showing transgressive-regressive relationship of Lodgepole and Mission Canyon Formations. Modified from Rose (1976).

Slightly later, the seas began transgressing into western Montana, depositing a black organic shale in many isolated, shallow marine basins. This shale is sometimes found at the base of the Lodgepole Formation. Shelf carbonate deposition was initiated with continued transgression into Montana and Wyoming during the middle to late Kinderhook, along with periodic influxes of terrigenous clastic sediments, and this makes up the thin-bedded limestones and shales of the Paine Member. By late Kinderhook time, the shelf margin had shifted to the east as the result of a general rise in sea level. During early Osagean time, as the seas continued to advance to the east, the influx of clastics ceased, resulting in the thin to thick-bedded limestones of the Woodhurst Member. The contact between these two members is best recognized at the type section of the Lodgepole Formation in the Little Rocky Mountains. It is often indistinct in other areas (Nordquist, 1953).

The seas reached their maximum extent by middle Osagean time. As they began regressing, the Mission Canyon Limestone was deposited as a massive to thick-bedded, almost pure limestone. Restricted circulation during the middle to late Osagean and again during early Meramecian time resulted in the widespread deposition of evaporites. These evaporitic zones are mainly present in the subsurface of central and eastern Montana, and are usually not present in outcrop. Severson (1952), Andrichuk (1955), and Middleton (1961), however, have shown that zones of solution breccia in the outcrop are correlative with zones of evaporites in the adjacent subsurface, and that they thus represent zones of the leached evaporites.

The Mission Canyon/Lodgepole contact is sometimes indistinct, and is often drawn arbitrarily (Holland, 1952; Norton, 1956). It is usually taken as the bottom of the first massive limestone unit (Laudon, 1955) (fig. 6), but this is certainly not the case everywhere, because the base of the Mission Canyon Formation is sometimes medium-bedded. This would result in the contact between the two formations being placed higher than it should be. In addition, the character of this contact may change rapidly over a short

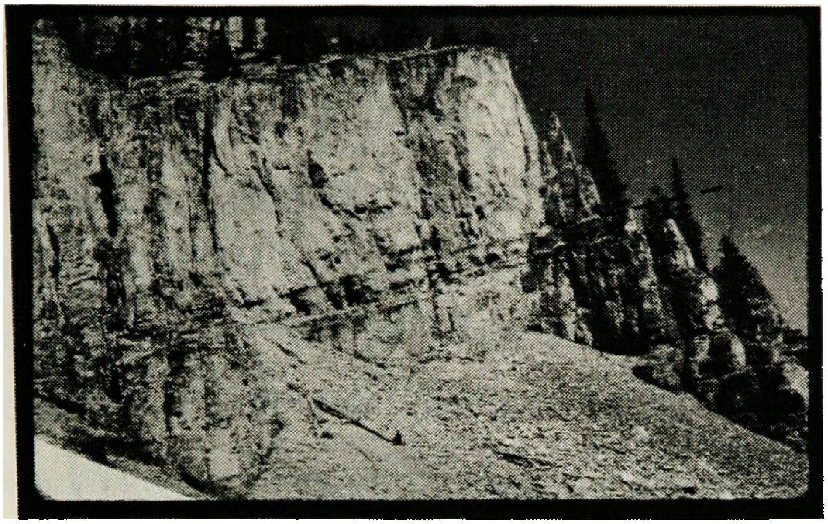


Figure 6. Mission Canyon/Lodgepole contact along Odell Creek in the Centennial Range of southwestern Montana. Note the thin beds of the Lodgepole Formation below the contact, and the massive beds of the Mission Canyon above.

distance. Moran (1971, p. 47), in his study of the Centennial Range of southwestern Montana, writes:

> "The contact between the Lodgepole and the overlying Mission Canyon is conformable and readily observable in the area. The contact is sharp, planar, and occurs between the very thin- and thin-bedded upper Woodhurst Limestone beds, and the very thick-bedded lower Mission Canyon Dolomite beds."

However Hall (1952, p. 93) who, in his study of the Mississippian in Montana and Canada, measured a section only 4 km from Moran's says:

> "The upper part of the Woodhurst Member is indistinguishable from the Mission Canyon Limestone..."

By late Meramec time, the seas had completely withdrawn to the west, and the Mission Canyon was subject to subaerial erosion and solution action. The seas returned to much of southwestern and central Montana via the Big Snowy Trough and Williston Basin, both of which were well defined at this time, and deposited the Big Snowy Group sediments in early to late Chester time. The rest of the state remained emergent until the deposition of the Amsden Formation sediments in the Pennsylvanian. These sediments collapsed and filled-in the caverns and channels of the karst topography caused by the extensive exposure. Both the karst topography developed at the top of the Mission Canyon and the varying amounts of time that the formation was exposed, have contributed to the extremely varying thickness of the Mission Canyon Formation. In the Bridger Range of western Montana the Mission Canyon Formation has been reported as from 290 m thick (McMannis, 1955) to as low as 130 m thick (Sloss and Hamblin, 1942). McMannis (1955) stated that the Mission Canyon is thickest wherever it is overlain by the Big Snowy Group, and thinnest wherever the Big Snowy Group is thin or absent. But near Three Forks, Montana, Robinson (1963) showed that the Mission Canyon varies considerably in thickness no matter how much of the Big Snowy Group is above it. Here the Mission Canyon ranges from 180 m to 210 m thick at Milligan Creek to 300 to 335 m thick just 3 km to the east (Robinson, 1963).

This hiatus in the middle to late Meramec is apparently represented throughout the state. The youngest fossils found in the Madison Group are corals and foraminifers of early Meramecian age (Sando and Mamet, 1974), and the oldest fossils overlying the post-Madison unconformity are foraminifers of middle Meramecian age (Mamet and Skipp, 1970; original reference not seen, see Sando and Mamet, 1974). Some geologists (for example Roberts, 1966) thought that all of the Mississippian Period was represented in the Williston Basin of eastern Montana (i.e., as the Madison Group, the Charles Formation, and the Big Snowy Group). But the youngest fossils from the Charles Formation are known to be early Meramecian in age (Sando, 1960), so a hiatus is probably also represented there.

Some workers have been able to subdivide the Mission Canyon Formation into two or three members in some areas. Denson and Morrisey (1952) defined two members of the Mission Canyon in the Big Horn and Wind River Basins of Wyoming; a 43 to 49 m thick lower member of bedded gray to tan finely crystalline cherty limestone and dolomite, and a 27 to 47 m thick upper member of massive bluish-gray limestone and finely crystalline dolomite, with the base marked by a breccia zone. In the Centennial Range of southwestern Montana, Moran (1971) divided the Mission Canyon into a 180 m thick lower member, and a 117 m thick upper member which may be correlative with the Charles Formation in the subsurface of eastern Montana. Sando (1972) examined the Madison Group in the Beartooth Mountains across the boundary between the Montana Province and the Wyoming Province. In each province he was able to identify three members: a lower cherty dolomite member deposited in a restricted marine environment, a middle "Cliffy Limestone Member" that has a 9 m thick breccia zone at its base (lower solution zone) which indicates deposition in a restricted lagoonal environment followed by deposition on shallow offshore marine banks, and an upper "Bull Ridge Member" with a 6 m thick breccia zone at its base (upper solution zone) which also indicates lagoonal deposition followed by marine bank deposition. The upper solution zone is approximately coincident with the Meramec-Osage boundary. Although these three units have not been traced out of the Beartooth Mountains, the lower solution zone has been traced as far away as the Big Snowy Mountains, and to the type section of the Madison Group at Logan, Montana (Smith and Gilmour, 1979).

History of Nomenclature

Peale (1893) originally defined the name "Madison Formation" for carbonate rocks exposed near Three Forks, Montana that were above the Three Forks Formation and below the Amsden Formation. He divided the formation into three divisions; from the bottom up they were "Laminated Limestones," "Massive Limestones," and "Jaspery Limestones." He gave no type locality, and apparently the name "Madison" did not imply that the type section was in the Madison Range nor along the Madison River, but simply was used because the Madison is one of the three rivers that join to form the Missouri River near Three Forks (Sloss and Hamblin, 1942). Iddings and Weed (1894) recognized similar rocks near Livingston, Montana, and Weed (1899), working in the Little Belt Mountains east of Helena, Montana, divided the Madison there into three members: the Castle Limestone, the Woodhurst Limestone, and the Paine Shale. Collier and Cathcart (1922) were the first to use the formal term "Madison Group" and named two formations: the Mission Canyon Formation, which is identical to the Castle Limestone of Weed (1899), and the Lodgepole Formation. Sloss and Hamblin (1942) synthesized Madison nomenclature and named the Paine and Woodhurst as members of the Lodgepole. They proposed the type section for the Madison Group to be along the Madison River near Three Forks. They also equated Peale's (1893) "Jaspery Limestones" with the Mission Canyon Formation, but others (for example Holland, 1952; Strickland, 1956) include the "Massive Limestones" in the Mission Canyon.

The Charles Formation was first defined by Seager (1942) as the carbonate, terrigenous, and evaporitic rocks at the base of the Big Snowy Group in the subsurface of the Williston Basin, and listed the type section as the California Company No. 4 Well in Petroleum County, Montana. Perry and Sloss (1943) agreed with Seager (1942), but Sloss (1950) and

Nordquist (1953) lowered the Charles to the upper part of the Madison Group. Towse (1957) and Roberts (1966) write that the base of the Charles can be considered to be the base of the lowest thick evaporite unit above the massive limestones of the Mission Canyon Formation. However, since the evaporites were deposited in different places at different times, other authors (for example Middleton, 1961; Sando, 1978) include some of the evaporites in the Mission Canyon Formation. Since the evaporites are present mainly in the subsurface, some workers believe that the Charles Formation should be restricted to the subsurface; others (Andrichuk, 1955; Roberts, 1966; Sando, 1967, 1974, 1976; Balster, 1971) have shown a correlation between some of the subsurface evaporite zones and breccia zones on surface outcrops that represent these evaporite zones removed by leaching. Sando and Dutro (1974) and Sando (1978) suggest that the Charles Formation should be considered as the carbonate-evaporite unit that overlies the Mission Canyon Formation in the subsurface of the Williston Basin, and that the name "Charles" should be confined to that area. This will be the useage followed in this paper.

LITHOLOGY

On the basis of fossil content, sedimentary and biogenic structures, thin-section petrography, and diagenetic modifications, seven carbonate lithotypes are recognized in the Mission Canyon Formation of southwestern Montana, as follows: (1) crinoidal biomicrite, (2) crinoidal oobiosparite, (3) biopelmicrite, (4) brachiopod biosparite, (5) fossiliferous micrite and sparse biomicrite, (6) dolomite, and (7) laminated limestone and dolomite.

Lithotype 1: Crinoidal Biomicrite (fig. 26).

Rocks of this lithotype consist of fossil fragments (usually crinoids) in a micrite matrix. Fossil fragments average 50% of the rock, with crinoids being by far the most abundant fossil. Occasionally brachiopods, peloids, ostracods, and bryozoans occur, but they usually amount to less than 5% of the rock, with forams, gastropods, and corals even more rare. The matrix consists of micrite which has been locally neomorphosed to microspar and, less frequently, pseudospar. Sparry calcite occurs infrequently in this lithotype, in amounts up to 10%, usually as cement filling-in "shelter" porosity, replacing fossils or cementing fossil molds, and as rim cement or syntaxial overgrowths on crinoid grains. Silica occurs occasionally in small amounts (less than 2%), either replacing fossils or as angular to subangular detrital quartz scattered throughout the rock. Partial dolomitization of the matrix is fairly common in this lithotype, usually characterized by scattered rhombic euhedra 40 to 60 microns in diameter, with some up to 125 microns. This lithotype generally has little or no porosity, except in a few instances where intercrystalline porosity exists in the partially dolomitized matrix. Pyrite may also exist in minor amounts, usually concentrated in burrows or in bioturbated areas.

This lithotype is present in all four sections at various horizons. It is the dominant lithology of the Camp Creek section.

Lithotype 2: Crinoidal Oobiosparite (fig. 17).

The rocks of this lithotype consist of fossil fragments in a matrix of sparry calcite cement. The fossil fragments average 70% of the rock, and consist mostly of ooids, often with crinoid particles as their nucleus, crinoids, and peloids, with lesser amounts of brachiopods, bryozoans, and forams. Some crinoid particles are heavily to completely micritized. All of these particles are usually well-rounded and well-sorted. The matrix of this lithotype consists entirely of sparry calcite cement. Many rocks of this lithotype are very tightly compacted, with microstylolitic contacts along grains and with cracked fossil fragments. These rocks are usually very well cemented and non-porous, except where porosity has been developed along fractures or stylolites.

This lithotype occurs in all sections except the one at Camp Creek. It is best developed in the Beartooth Mountains and in the Snowcrest Range, where it is present at various horizons in thin to thick beds from 1 to 5 m thick. In the Centennial Range it occurs only once, as a 2 m thick bed at the bottom of the section.

Lithotype 3: Biopelmicrite (plate 1B).

This lithotype consists of rocks made up of fossil fragments in a matrix of peloids, peloidal micrite, and microspar. The fossil fragments make up about 20% of the rock, with crinoids, brachiopods, and bryozoans the most abundant fossils. The matrix consists of small, lumpy, dark brown, well-rounded and well-sorted peloids, averaging about 50 to 75 microns in diameter. About half of this matrix has neomorphosed to microspar. Sparry calcite exists in amounts less than 5%, replacing or cementing shell fragments, and as pseudospar. Silica occurs in trace amounts, as subangular detrital quartz silt and as authigenic silica replacing some shell fragments.

This lithotype is present only in the Snowcrest Range as a 6 m thick, thin to medium-bedded cliff-forming unit near the bottom of the section. It is also present as a very thin lense in the middle of the section.

Lithotype 4: Brachiopod Biosparite (fig. 19C).

Rocks of this lithotype consist of brachiopod fragments in a matrix of micrite, microspar, and pseudospar, with sparry calcite cement. Brachiopods make up about 35% of the rock, with crinoids, ostracods, bryozoans, and corals much less abundant. Sparry calcite exists in amounts up to 50%, as cement and as neomorphic(?) overgrowths on crinoid grains. A slight amount of compaction has resulted in occasional cracked fossil fragments.

This lithotype occurs only in the Snowcrest Range section as a 14 m thick, thin to medium-bedded slope and cliff-forming unit immediately above the 6 m thick interval of Lithotype 3.

Lithotype 5: Fossiliferous Micrite and Sparse Biomicrite (plate 1F).

This lithotype is made up of rocks with a fine-grained carbonate mud matrix with minor amounts of fossil grains, sparite, and dolomite. The matrix is mostly of micrite, in places neomorphosed to microspar and pseudospar, and makes up from 85% to over 99% of the rock. The grains consist of small fragments of crinoids, brachiopods, ostracods, bryozoans, and other unidentified fragments. Sparry calcite cement occurs in amounts up to 5%, usually as cement in molds or vugs. Occasionally, birdseye structures are also present. In many of the rocks of this lithotype, the mud is slightly dolomitized, usually with scattered dolomite euhedra 40 to 60 microns in diameter.

This lithotype is found only in the Snowcrest Range section and in the Beartooth Mountains section. It occurs at various horizons as thin to massive, slope and cliff-forming units.

Lithotype 6: Dolomite (fig. 25).

Rocks of this lithotype are made up of dolomite crystals ranging in size from 10 to 1000 microns, with about 75% of them in the range of fine to medium crystalline (16 to 250 microns; terminology from Folk, 1959), and most of those between 30 and 80 microns. This lithotype is divided into four subtypes based on grain size.

Subtype 1: Very fine-grained dolomite. The dolomite grains in this subtype average less than 16 microns in diameter, and they are usually subhedral. These rocks often have up to 2% porosity produced by vugs 10 to 25 microns in diameter. Sparry calcite may also exist, in amounts up to 2%, usually as cement-filled vugs and crinoid molds. Both void space and sparry calcite cement can exist in the same rock.

Subtype 2: Fine-grained dolomite. The dolomite grains in this subtype average from 16 to 62 microns in diameter, and they are usually euhedral to subhedral. Pyrite and detrital quartz occur in trace amounts in this subtype, usually scattered throughout the rock. Sparry calcite cement occurs frequently in amounts up to 15%, usually as cement-filled vugs and molds. These rocks often have from 5 to 10% porosity produced by vugs, molds, and solution-enlarged molds of crinoids. In addition there are often large vugs up to 8 mm in diameter of unknown origin.

Subtype 3: Medium-grained dolomite. The dolomite grains in this subtype average from 62 to 250 microns in diameter, and are usually euhedral to subhedral. Most of these rocks have a considerable amount of porosity, frequently as much as 20%, and usually as intercrystalline void space and as solution-enlarged molds of crinoids. Dolomitized crinoids and scattered patches of sparry calcite cement also occur occasionally.

Subtype 4: Coarse-grained dolomite. The dolomite grains in this subtype average from 250 to 1000 microns in diameter, and are usually euhedral to subhedral. In thin-section they are usually featureless, with occasional vuggy porosity. This subtype is the least common of the four.

Dolomite occurs in all sections except the Camp Creek section. In the Centennial Range the entire section is massive dolomite except for the lowermost 8 m, which is limestone. The sections in the Snowcrest Range and in the Beartooth Mountains each contain about 40% dolomite. Further discussion of the stratigraphic position of dolomite will be deferred until later in the text.

Lithotype 7: Laminated Limestone and Dolomite (plate 1A).

More than 70% of all laminated rocks observed in this study are very fine-grained to medium-grained dolomites. Laminations in dolomites are usually produced by alternating layers of coarse and fine-grained dolomite crystals. In addition, they are frequently mottled by patches and horizontally-oriented lenses of coarse-grained dolomite crystals. Laminated dolomites usually lack fossils, and often have birdseye and mudcrack features.

Laminated limestones are of two general types: those whose laminations are produced by alternating layers of coarse and fine-grained particles, and those whose laminations are produced by algae.

Laminated rocks are found in all sections except the Camp Creek section. They occur throughout each section at various stratigraphic intervals.

BRECCIATION

Two types of breccias can be recognized in the Mission Canyon Formation of the studied area: karst breccias and evaporite-solution breccias. Roberts (1966) has a good discussion of both types in the Madison Group near Livingston, Montana.

Karst Breccias

Karst breccias were developed, starting in late Meramec time, when the seas withdrew to the west from Montana. The resulting period of erosion developed a karst topography on top of the Mission Canyon Formation. The karst breccias were formed by the collapse of sinkholes and cavern roofs prior to or during the deposition of the overlying rocks (either the Mississippian Big Snowy Group or the Pennsylvanian Amsden Formation). These breccias have a sharp, well-defined lower boundary and a gradational, poorly-defined upper boundary. They are made up of varying sizes of angular to subangular limestone and/or dolomite clasts, cemented by sparry calcite. Most of the clasts range from 1 mm to 100 mm in size, but some are so large (up to 2 m in diameter) that they are difficult to identify as clasts. They are sometimes red-colored from the weathering effect of the overlying sediments. In Montana, they have been described near Livingston, where there are clastic dikes of red Amsden Sandstone that fill a solution channel that is 60 m deep at the top of the Mission Canyon Formation (Sloss and Hamblin, 1942), near Toston where there is up to 30 m of relief on the top of the Mission Canyon Formation (Robinson, 1963), near Riceville (Walton, 1946), and in other areas throughout the state. Henbest (1958) describes similar features in the Mississippian of Wyoming, southwestern Colorado, and northwestern New Mexico.

Although these breccias occur sporadically and are not laterally continuous, they are usually evident wherever the contact between the Mission Canyon Formation and the overlying formation is visible. This contact, however, is often poorly exposed in the area of this study. Of the four sections measured, a unit interpreted as a karst breccia was observed at the top of the section in the Centennial Range and in the section along Camp Creek. In neither section is the contact of the Mission Canyon Formation with the overlying Amsden Formation exposed. The breccias exist as scattered pods from 20 cm in diameter up to lenses 1 m thick and 5 m long. They are often red-colored in the Centennial Range. In the Beartooth Mountains, a karst breccia was not noted, but some large solution caves were observed with binoculars on the side of an inaccessible cliff near the top of the section that may be a remnant of the karst topography. No karst breccia was observed in the Snowcrest Range.

Evaporite-Solution Breccias

Evaporite-solution breccias were formed when evaporite layers in the Mission Canyon Formation were leached by groundwater as a result of surface or near-surface exposure (fig. 7). These breccias probably formed later than the karst breccias because they are found only in areas of Late Cretaceous and Early Tertiary uplift (Roberts, 1966), although Middleton (1961) suggested that at least some solution brecciation may be contemporaneous with karst brecciation.

A major solution breccia zone within the Mission Canyon Formation, which is sometimes used as a key bed for subdivision of the formation, is present throughout much of western Montana and is the approximate boundary between the

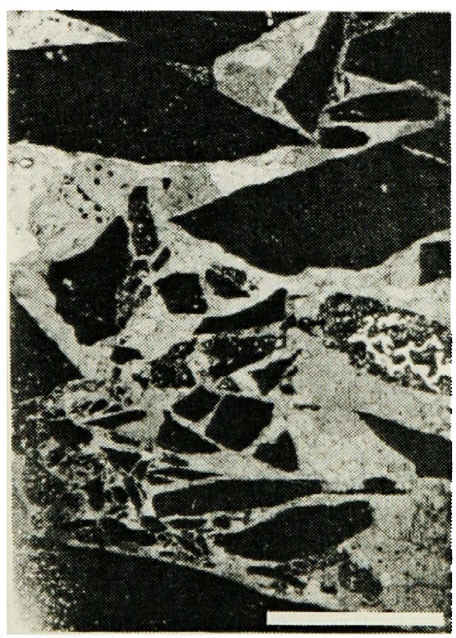


Figure 7. Evaporite-solution breccia. Dark angular clasts are very finely-crystalline dolomite. Lighter areas are sparry calcite cement. Thin section, plane light. Centennial Range, U.M.P. #7004. Bar=10 mm.

Osage and the Meramec (Laudon, 1948; Roberts, 1966). At one time, this breccia was taken to represent an unconformity within the Mission Canyon Formation (Berry, 1943; Laudon, 1948; Denson and Morrisey, 1952; Strickland, 1956), but later workers (Norton, 1956; Middleton, 1961) have shown this interval to be a breccia resulting from the solution of evaporite beds, and that it thus does not represent an unconformity.

Solution breccias are usually not difficult to distinguish from karst breccias. Although both types usually have distinct lower contacts and indistinct upper contacts, evaporite solution breccias are usually much more laterally continuous than those attributable to karst; often extending for a distance of tens of kilometers or more. Additionally, solution breccias are usually associated with shallow water sequences and/or penecontemporaneous supratidal dolomitization, and may have anhydrite-replacement pseudomorphs (Beales and Oldershaw, 1969). Solution breccias tend to have clasts with small sizes, unsorted at the base, and grade upward to slightly fractured roof rocks at the top (Severson, 1952). Karst breccias, on the other hand, tend to have more matrix and often fill cavities that widen upward (Roberts, 1966), and thus are more likely to be vertically oriented rather than horizontally. Roberts (1966) also reported some success in differentiating between the two types of breccias on the basis of the clay mineralogy in the insoluble residues; with kaolinite predominating in the karst breccias and illite in the solution breccias.

Evaporite solution breccias occur in all sections except the Camp Creek section. They are most abundant in the Centennial Range, with intercalated breccia zones ranging from 1 to 17 m thick, occurring intermittently from the bottom of the section to near the top. In the sections in the Beartooth Mountains and in the Snowcrest Range they are much less prevalent. In the Snowcrest Range there are two or three zones of brecciation 1 to 9 m thick at the bottom and near the center of the section, and in the Beartooth Mountains there are two zones 1½ to 3 m thick; one at the bottom and one near the top of the section.

POROSITY

The origin and evolution of porosity in carbonate rocks differs in many ways from that in detrital rocks. Primary depositional porosity in sandstones, for example, averages from 25 to 40%, while the porosity in most ancient sandstones is from 15 to 30% (Choquette and Pray, 1970). Porosity reduction is effected by compaction and cementation, and usually the final porosity of the lithified rock is simply somewhat-reduced primary interparticle porosity. By comparison, primary depositional porosity in carbonates averages 40 to 70% (Bathurst, 1966) and yet porosity in ancient limestones is usually less than 5%. Most or all of the porosity reduction in carbonates comes from cementation of the pore space, with possibly a smaller amount of reduction coming from compaction, even though many carbonates show little or no evidence of compaction (Pray, 1960; Zankl, 1969). Compaction may be reduced or prevented by the development of early cements, especially if a substantial framework, organic or otherwise, exists. Early diagenetic cements, however, are sometimes involved in the compaction process (Moore et al., 1981). Generally limestones composed of competent grains, such as ooids, can retain their porosity under compressive stress better than those composed of less competent grains, such as pellets (Brock, 1980).

Primary interparticle porosity can, however, be preserved in limestone, although it is infrequently encountered. It is most common in arid climates, where the lack of fresh water prevented a significant amount of cementation (Longman, 1981), and if cementation in the marine phreatic or vadose zone did not occur or was of minor consequence. Cementation and compaction are thus much more effective in reducing porosity in limestone than they are in sandstone.

Later diagenetic events are usually necessary in order to effect porosity in carbonates. Diagenetic events that can produce pore space include dissolution of cement, mud, or fossils to produce molds or vugs; fracturing, including brecciation; and dolomitization. Excessive lithostatic pressure as a result of deep burial can also produce secondary porosity (Donath et al., 1980; Moore et al., 1981). These secondary pores then could become cemented by another generation of cement, and it is not unusual for these recemented pores to become pore space again at some later time, through other diagenetic events. Longman (1980) suggests that it might be most useful to think of a tightly cemented, non-porous carbonate rock as the normal end product of diagenesis, and that any rock with significant porosity is simply at an interrupted stage of its history.

Porosity in the limestones and dolomites of this study take the following forms (see Choquette and Pray, 1970, for terminology):

1. Vugs of unknown origin.
2. Molds, usually of crinoids.
3. Solution-enlarged molds, usually of crinoids.
4. Fracture, including stylolitic.
5. Intercrystalline pores.
6. Micropores.

All of these types of porosity are secondary in origin except possibly for some microporosity. The molds and solution-enlarged molds are fabric selective, while the stylolites, fractures and most of the vugs are not. The presence of microporosity is difficult to evaluate. Micropores occur within coatings on grains such as ooids (Illing, 1954), in micritic envelopes, in carbonate mud, and as small voids left over when aragonite inverts to calcite (Pittman, 1971). Because of their small size (less than 1 micron) they are impossible to see using normal petrographic microscope techniques. In addition, the normal 30 micron thickness of a thin section almost invariably results in the amount of porosity being underestimated, often by more than 100% (Halley, 1978), and of course pores smaller than 30 microns cannot be seen at all. To avoid these difficulties, porosity volumes were estimated, wherever possible, from acetate peels, polished slabs using a binocular microscope, and by using ultra-thin (5 to 10 microns thick) thin sections (see Lindholm et al., 1973).

In the rocks examined for this study, porosity was found to be almost always associated, in some way, with dolomite. Only about 10% of the rocks with any significant amount of porosity are limestone; all the rest are dolomites or calcareous dolomites. The most commonly observed type of porosity is a dolomitic rock with molds, solution-enlarged molds, or vugs, with or without intercrystalline porosity. Almost invariably, the shape of the molds suggests that they were originally crinoids (fig. 8), and in most cases the solution-enlarged molds are most likely enlarged crinoid molds. Careful observation of the solution-enlarged molds will usually yield at least one that can be recognized as crinoid-shaped, and it is probably safe to assume that other vugs or solution-enlarged molds of similar size and shape were originally crinoids. These molds and solution-enlarged molds average .75 to 1 mm in size, with some as large as 4 to 5 mm. There are usually also smaller vugs of .1 to .2 mm in size that are of unknown origin. These smaller vugs can make up from 1 to 5% porosity in the rock. It is difficult to determine if these smaller vugs

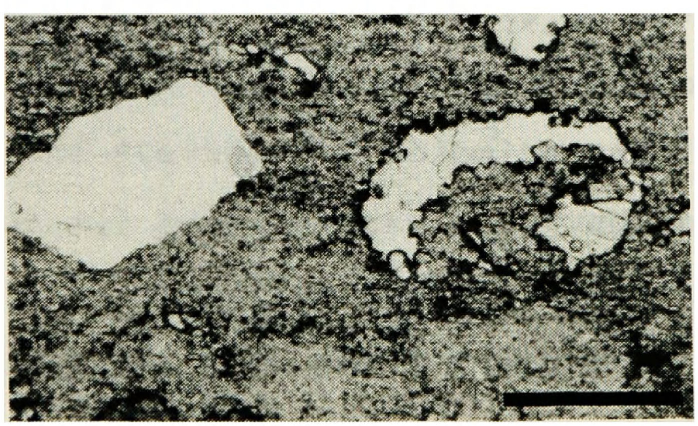


Figure 8. Crinoid mold (left), and one cemented with sparry calcite. Acetate peel, plane light. Centennial Range, U.M.P. #7010. Bar=.8 mm.

are fabric selective or not. In addition, about 40% of the dolomites with moldic porosity will also exhibit intercrystalline porosity. These pores take the form of irregularly shaped spaces between the crystals of dolomite. These pores are usually smaller than 60 microns, and are observed right down to the limit of resolution with the microscope. Because they can be so small, this is one type of porosity that is likely to be underestimated by visual observation or point counting. The best way to determine the actual amount of porosity in this case would be by some physical method, such as capillary pressure measurement (Wardlaw, 1976).

The next most common type of porosity observed is of a dolomitic rock with intercrystalline porosity only. In all cases these rocks have a matrix made up of medium crystalline to very finely crystalline dolomite. No particles exist in these rocks, and there is no evidence of any replaced particles, such as dolomitic pseudomorphs of crinoids. These rocks average 84% dolomite and 12% porosity.

About 85% of all porous rocks observed in this study are of the two types described above. Other types of porosity include limestone with small (less than 100 microns) vugs of unknown origin, fracture porosity, which is quantitatively most important in solution breccias where calcite cement did not completely fill fractures in the rock, and stylolitic porosity, where stylolites are open or where insoluble clay residues have been removed along the stylolite (fig. 9). Because most of the porosity is associated with dolomite, a discussion of the origin of porosity will be deferred to the section on dolomitization.

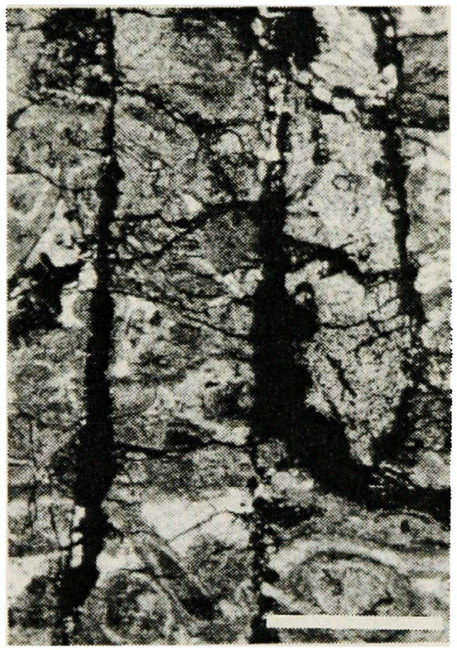


Figure 9. Porosity (black) developed along a stylolite. Thin section, crossed nicols. Beartooth Mountains, U.M.P. #6982. Bar=.8 mm.

DIAGENESIS

Rocks of the Mission Canyon Formation in southwestern Montana provide a variety of diagenetic "facies" that lend themselves to interpretation of the geologic history of the formation. In this section, the specific types of diagenesis that have occurred in the rocks under study will be discussed. For the purpose of this report, diagenesis is defined following the definition proposed by Larsen and Chilingar (1979, p. 393 and 395) and it "includes all physicochemical, biochemical and physical processes modifying sediments between deposition and metamorphism." Excellent discussions of carbonate diagenesis are available from many authors, in particular Bathurst (1958, 1975), Murray (1960, 1964), Lucia (1962), Folk (1965, 1974), Bricker (1971), Larsen and Chilingar (1979), Moore (1979), Longman (1980), and Flugel (1982).

The diagenetic changes noted in the rocks of this study are as follows: (1) inversion of aragonite to calcite, (2) micritization, (3) neomorphism, (4) compaction and pressure-solution, (5) silicification, (6) pyritization, (7) cementation, and (8) dolomitization.

Inversion of Aragonite to Calcite

In modern carbonate sediments, aragonite and high Mg-calcite make up approximately 70% of the sediment (Stehli and Hower, 1961), while in ancient carbonate rocks the mineralogy is usually entirely low Mg-calcite (or dolomite). Bathurst (1964) and Friedman (1964) show two general ways in which originally aragonite particles can become calcite:

1. Replacement.
2. Solution-deposition.

These two processes differ only in the size of the void space that exists during the process. Dodd (1966) enlarged upon this scheme and described four ways in which solution-deposition can occur.

A bryozoan fragment that has been replaced by calcite is shown in figure 10.

Figure 10. Bryozoan fragment that has been replaced by calcite. The matrix of this rock is made up of a considerable amount of microspar. Thin section, plane light. Snowcrest Range, U.M.P. #7030. Bar=.8 mm.

Micritization

Micritization of carbonate particles is caused by endolithic algae or fungi which bore into the particle and then live in the bored holes (Bathurst, 1966). After the algae or fungi die, the vacated hole usually gets filled-in with micrite (fig. 11). Repeated boring can cause the entire particle to become replaced by micrite (fig. 12). If this is the case, the particle may appear as a nondescript "lump" of dark brown micrite, and would then be termed a "peloid."

In addition to the centripetal replacement just described, micritic coatings can occur as centrifugal accretions, i.e., as a constructive envelope (Kobluk and Risk, 1977). In neither case will these envelopes form

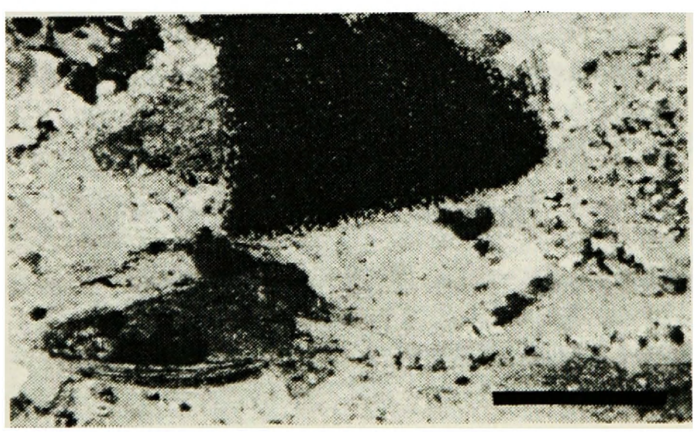


Figure 11. Crinoid grain (black) with a bored and micritized rim. Note the single-crystal extinction characteristic of crinoid particles. Thin section, crossed nichols. Snowcrest Range, U.M.P. #7051. Bar=.8 mm.

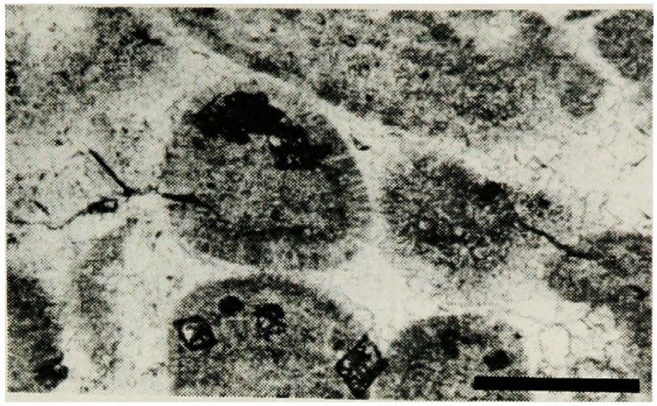


Figure 12. Oosparite/grainstone. Particles in this rock are nearly completely micritized. This photo shows the poorly preserved radial structure of a micritized ooid (center). Thin section, plane light. Beartooth Mountains, U.M.P. #6996. Bar=.2 mm.

under agitated conditions (Kobluk and Kahle, 1978), and they are often very difficult to tell apart.

In the Mission Canyon Limestones under study, micritization, where it occurred, took several forms: (1) as thin micritic envelopes (fig. 13), (2) as thick micritic coatings, and (3) as completely or almost completely micritized particles (fig. 12).

A micritic envelope around a fossil fragment can remain after the fragment itself has been dissolved away. It will then preserve the outline of the fragment and may help to identify it.

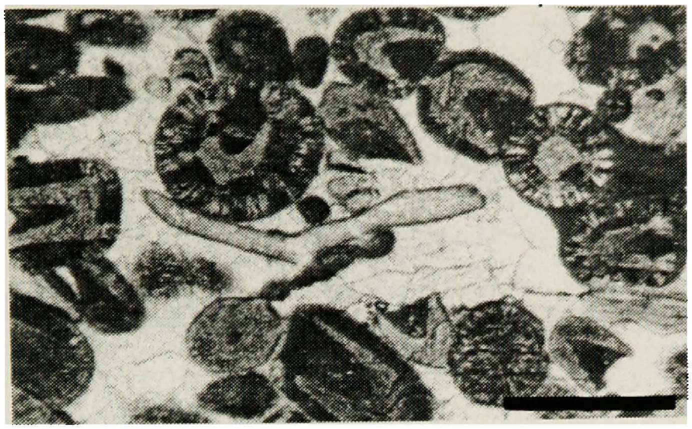


Figure 13. Shell fragment with a micritic envelope fractured by compaction. The fragment has been dissolved and cemented or replaced by sparry calcite. Thin section, plane light. Beartooth Mountains, U.M.P. #6987. Bar=.8 mm.

Neomorphism

The term "neomorphism" was introduced by Folk (1965, p. 21) as "a comprehensive term of ignorance...for all transformations between one mineral and itself or a polymorph..." It thus includes both inversion and recrystallization, and is used whenever the composition of the original starting material is unknown. As it is usually impossible to determine the original composition of any particular ancient calcareous mud or the time at which the transformation in question took place (i.e., whether before or after an aragonitic mud had inverted to calcite), "neomorphism" is the proper term to use for processes involving carbonate mud.

Three different neomorphic processes have affected the rocks examined for this study:

A. The neomorphism of aragonite or high Mg-calcite mud to micrite. In any limestone that originally contained carbonate mud, its neomorphism to normal 1 to 4 micron micrite is an important diagenetic process (Folk, 1965). Both Folk (1965) and Bathurst (1975) have excellent discussions of this process.

B. The neomorphism of micrite to microspar. Microspar is the product of aggrading neomorphism of micrite to larger sized particles of approximately 4 to 30 microns (Folk, 1965). Microspar can sometimes be difficult to distinguish from similar size crystals of pore-filling calcite cement. Bathurst (1975) discusses several ways to distinguish between the two. Additionally, care must be taken to avoid confusing microspar with vadose silt (see Folk (1965) and Bathurst (1975)).

In the Mission Canyon Formation of southwestern Montana, microspar is common in all sections except the dolomitized section in the Centennial Range. It usually occurs as scattered patches up to 1 mm in diameter as the apparent result of the local coarsening of micrite. The section along Camp Creek contains an abundance of microspar (fig. 14). Deformed crinoids are found in a matrix of microspar which has, to a large extent, almost completely obliterated the original texture of the rock.

C. The neomorphism of micrite or microspar to pseudospar. If microspar continues to grow larger than an arbitrarily

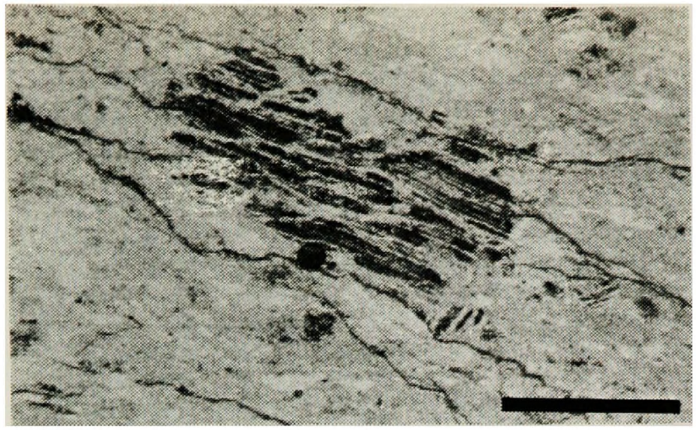


Figure 14. "Stretched" crinoid in a crinoidal biomicrite/packstone. There is a considerable amount of microspar in the matrix of this rock. The crinoid grain was deformed by folding. Notice the linear distribution of micrite (lighter color) along the glide planes in the crinoid. This is the "primary recrystallization" of Bathurst (1975) or the "degrading recrystallization" of Folk (1965). Thin section, crossed nicols. Camp Creek, U.M.P. #7057. Bar=.8 mm.

set upper limit of about 30 microns, it is then called pseudospar. Pseudospar is relatively uncommon in the rocks of the Mission Canyon Formation under study, occurring sporadically as occasional larger grains within a mass of microspar, or as syntaxial overgrowths on crinoid particles. In figure 15, there are several lines of evidence that show that the overgrowth on the crinoid grain is a product of aggrading neomorphism and is not pore-filling cement. Although Folk (1965) suggests that rim cement on a crinoid grain can push aside micrite as it grows, Lucia (1962) and Evamy and Shearman (1965, 1969) have shown that any significant amount of lime mud in contact with a crinoid grain prevented the development of rim cement. Bathurst (1975)

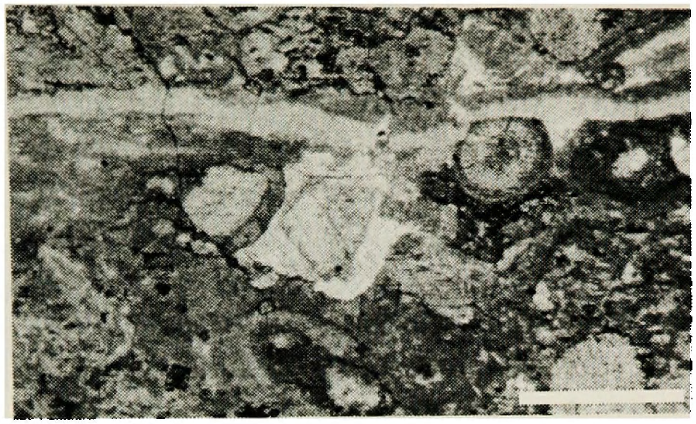


Figure 15. Syntaxial overgrowth on a crinoid grain. This is not pore-filling sparry cement. Although in some instances sparry cement overgrowths may be able to force the matrix aside without replacing it (Folk, 1965), in this example the cloudy nature of the spar and the highly irregular outer boundary, especially along the top edge, is persuasive evidence that the spar is of a neomorphic origin. Thin section, plane light. Beartooth Mountains, U.M.P. #6993. Bar=.8 mm.

also claimed that in a mudstone or a wackestone (i.e., in a mud-supported rock), any overgrowth on a crinoid is necessarily a syntaxial overgrowth. If this is so, then in the rock shown in figure 15, the overgrowth must be a syntaxial overgrowth, assuming that the grains and the mud were deposited simultaneously. If the grains were deposited first, then an episode of rim cementation could occur that would be terminated whenever the carbonate mud was washed in. Unfortunately, in this example, there is not enough evidence to show whether the mud was deposited along with the grains or whether it came later. There is, however, textural evidence to suggest that the overgrowth is of the neomorphic type. The evidence consists of the cloudy nature of the spar in thin-section, which is the result of minor particles or "ghosts" of micrite that were not completely neomorphosed to pseudospar, and of the jagged outline of the sparite crystal, especially along its top edge, with large patches of micrite "floating" in it. The sparite in this example is thus properly termed a "syntaxial overgrowth" (Bathurst, 1958).

Compaction and Pressure-Solution

Compaction effects in the rocks of this study range from no visible signs of compaction to minor grain interpenetration to stylolites with amplitudes of up to 15 mm. In general, however, most of the rocks show evidence of little or no compaction. In figure 16, judging by visual

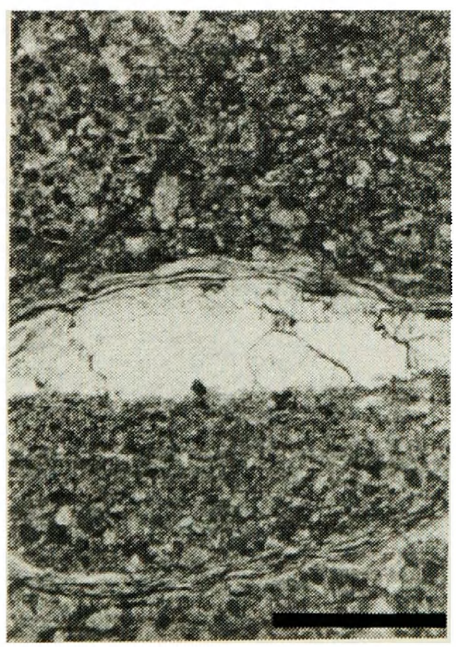


Figure 16. Floored cavity within a brachiopod shell. Area occupied by sparry cement was once primary porosity. The micritic sediment within the shell is virtually identical to that outside of the shell. This is a good indication of the negligible amount of compaction that took place. Thin section, plane light. Beartooth Mountains, U.M.P. #6986. Bar=.8 mm.

observation of the micrite inside the shell compared to outside of it, it is clear that the sediment underwent little if any compaction. On the other hand, one might argue that just because there are no obvious signs of compaction, such as broken fossil grains, that the sediment was not compacted. Shinn and others (1977) have conducted experiments where fine-grained limestones were compacted 50 to 65% under a pressure of 550 kg/cm^2, and fossil fragments within were not broken. Brown (1969) compacted mud-supported sediments under loads of more than 1000 kg/cm^2 and only rarely did any shells break. Unbroken fossils, therefore, are not evidence that a sediment has not compacted, at least in a mudstone or wackestone.

Meyers (1980) showed that compaction could reduce intergranular volumes in a muddy sediment by up to 50%, and that compaction was essentially complete by the time a sediment had been buried to a depth of 2000 m. This data is consistent with the findings of Fruth and others (1966) whose experimental work showed that, even with varying initial porosities, all sediments studied had approximately 35% porosity under compaction equivalent to burial at 1100 m, 22 to 28% at 4600 m, and 14 to 25% at 9100 m.

Rocks of the Mission Canyon Formation examined during the course of this study exhibit three types of compaction features: (1) minor grain interpenetration and pressure-solution, (2) "fitted" fabrics and solution seams, and (3) stylolites.

Minor grain interpenetration and pressure-solution are best shown in oolitic rocks, because the original shape of a partially dissolved ooid is easily determined. In figure 17, it is easy to tell how much any particular ooid has penetrated its neighbor. Note especially the two ooids at the bottom center of the photograph. This type of fabric is also found between crinoid particles in this rock and in other rocks.

Fitted fabrics and solution seams are not common in the rocks of this study. Where they exist, they consist of strongly fitted-together particles with solution seams along the particle contacts. Apparently these solution seams are

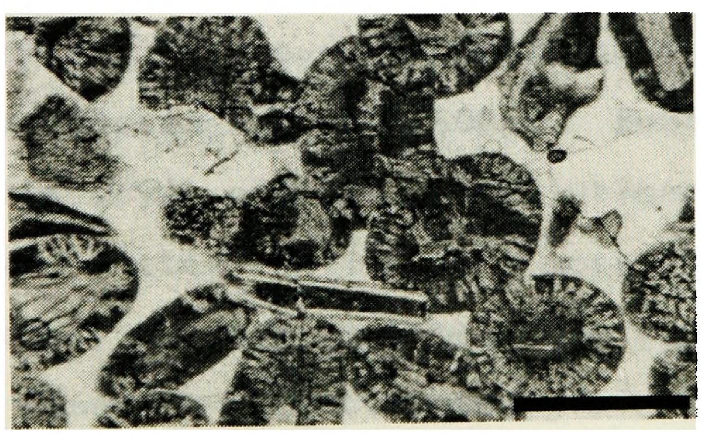


Figure 17. Oosparite/grainstone. The nuclei of many of these ooids are crinoid fragments. The linear grain near the center has been cracked by compaction with the crack later healed by sparry calcite cement. Many ooids are in microstylolitic contact with one another. Thin section, plane light. Beartooth Mountains, U.M.P. #6987. Bar=.8 mm.

the result of solution transfer where, when grains are dissolved at their points of contact with other grains, the solution is precipitated as cement somewhere else (either on that same grain or elsewhere), and the result is a particle orientation with the long axes arranged horizontally (Bathurst, 1975).

Stylolites occur frequently in the rocks of this study. They are generally sutured, and less frequently occur as a simple wave or sharp peak (terminology from Park and Schot, 1968). Amplitudes range up to 15 mm, and almost all are parallel to the plane of bedding. Most stylolites contain insoluble residues of clay or other material; some apparently have had this clay removed, resulting in porosity along the stylolite (fig. 9), and some contain dead oil or bituminous compounds.

Dunnington (1954) estimated that stylolites would not form until a rock had been buried to at least 600 m, while Scholle (1971) estimated at least 500 m. Buxton and Sibley (1981) related the style of pressure-solution to cementation. Solution seams and fitted fabrics were most common in packstones, wackestones, and poorly-cemented grainstones, while stylolites were found in well-cemented grainstones. Their observations do not seem to agree with those of this study. Although there are not enough grainstones in the sections studied to make any statistically significant correlations with the results obtained by Buxton and Sibley (1981), most stylolites observed in the rocks under study occur in packstones and in dolomites.

Silicification

Silica is moderately common in the rocks of this study and occurs in all four sections. Authigenic silica usually occurs as a light gray to cream colored chert, in beds up to 5 cm thick, stringers 2 to 5 cm thick, and nodules up to 10 cm thick. The nodules are usually irregularly shaped, and are flattened in the plane of bedding. The chert usually replaces both the particles and the matrix of the rock, although in thin-section, silica can be seen that selectively replaces only fossil fragments. Crinoids are the fossil most likely to be silicified, with brachiopods and gastropods somewhat less-likely candidates. This observation does not seem to agree with the results obtained by Newell and others (1953), who found bryozoans, brachiopods, and mollusks more susceptible to silicification than crinoids.

The crystal size of the authigenic silica ranges from cryptocrystalline to about 15 microns. In some cases where the rocks are dolomitized, it can be seen that silicification both preceded the dolomitization process and prevented it from dolomitizing the entire rock (fig. 18A).

Detrital quartz silt also occurs in the rocks of this study. It is present in all four sections, although it is very rare in the Camp Creek section. Where present, it

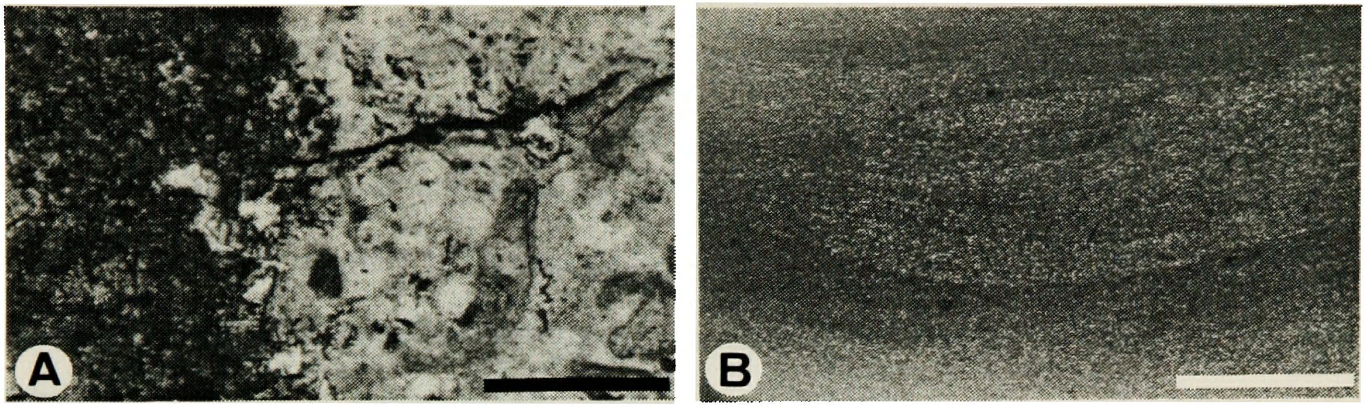


Figure 18. Common silicification textures. A. Portion of a chert nodule in medium-crystalline dolomite. The preservation of the original texture in the chert (light color) is better than in the dolomite, indicating that silicification preceded dolomitization. Thin section, plane light. Snowcrest Range, U.M.P. #7043. Bar=.8 mm. B. Cross-bedded detrital quartz silt (white spots) in finely-crystalline dolomite. Thin section, plane light. Beartooth Mountains, U.M.P. #6976. Bar=10 mm.

usually occurs in amounts less than 2%, as subangular quartz silt scattered throughout the rock. No attempt was made to analyze the sources nor the mechanics of distribution of detrital constituents of these rocks.

Pyritization

Pyrite is relatively uncommon in the rocks of this study. It occurs in all sections except the Camp Creek section. It usually exists in trace amounts, as 10 to 30 micron euhedra, often concentrated in burrows or in bioturbated areas of the rock. It also almost invariably occurs with chert. It is interpreted to be authigenic on the basis of its euhedral form.

The specific conditions under which the pyrite was formed could not be determined, although its association with burrowed and bioturbated sediments suggests a bacterial origin under euxenic conditions (Larsen and Chilingar, 1979). Euxenic conditions can exist in shallow water below the sediment/water interface (Moretti, 1957). Here, in the presence of organic matter, anaerobic bacteria can reduce sulfate in sea water to yield H_2S, which can combine with iron to form pyrite. In carbonate sediments the limiting factor is usually the small amount of iron available (Berner, 1970). Both pyrite and calcite are stable at a pH of 8.0 and an Eh of -0.3 (Krumbein and Garrels, 1952).

<u>Cementation</u>

Three types of cement are found in the rocks of the Mission Canyon Formation under study: (1) radial fibrous cement, (2) coarse, blocky sparite, and (3) rim cement on crinoids. Micrite cement and syntaxial overgrowths on crinoid grains were previously discussed under neomorphism (p. 39). It should be re-emphasized here that all sparry calcite found in these rocks is not necessarily cement; some of it is of a neomorphic origin. Fabric criteria were previously discussed and/or referenced for neomorphic spar, and Bathurst (1975) discusses criteria for the recognition of cement.

Radial fibrous cement exists as isopachous crusts up to 20 microns thick surrounding shell fragments. It occurs frequently in grainstones. In figure 17, the brachiopod spine(?) in the center of the photo has a thin coating of radial fibrous cement all along its outer perimeter. Note that, particularly along its left end, this cement has prevented the spine from coming in contact with the ooids, and that the ooids are in microstylolitic contact with one another, and have no layer of fringing cement.

Figure 19 shows several good examples of the texture of pore-filling, coarse, blocky spar. The ostracod shell in figure 19A shows the typical fabric of pore-filling sparry cement particularly well. Note especially the numer-crystals along the shell wall and how they coarsen and become less numerous toward the center. This is caused by competetive growth which favors one particular crystal orientation over another (Bathurst, 1975) and possibly because, as the pore space fills up with cement, the diffusion rate of the water moving through the pores decreases, permitting the crystals to grow larger (Longman, 1980). Coarse, blocky spar is particularly well developed in oolitic and crinoidal grainstones, and also wherever it fills-in original shelter porosity in packstones (fig. 19C). Coarse, blocky spar also occurs frequently filling-in secondarily derived pore space, especially the vugs, molds, and solution-enlarged molds of crinoids developed during

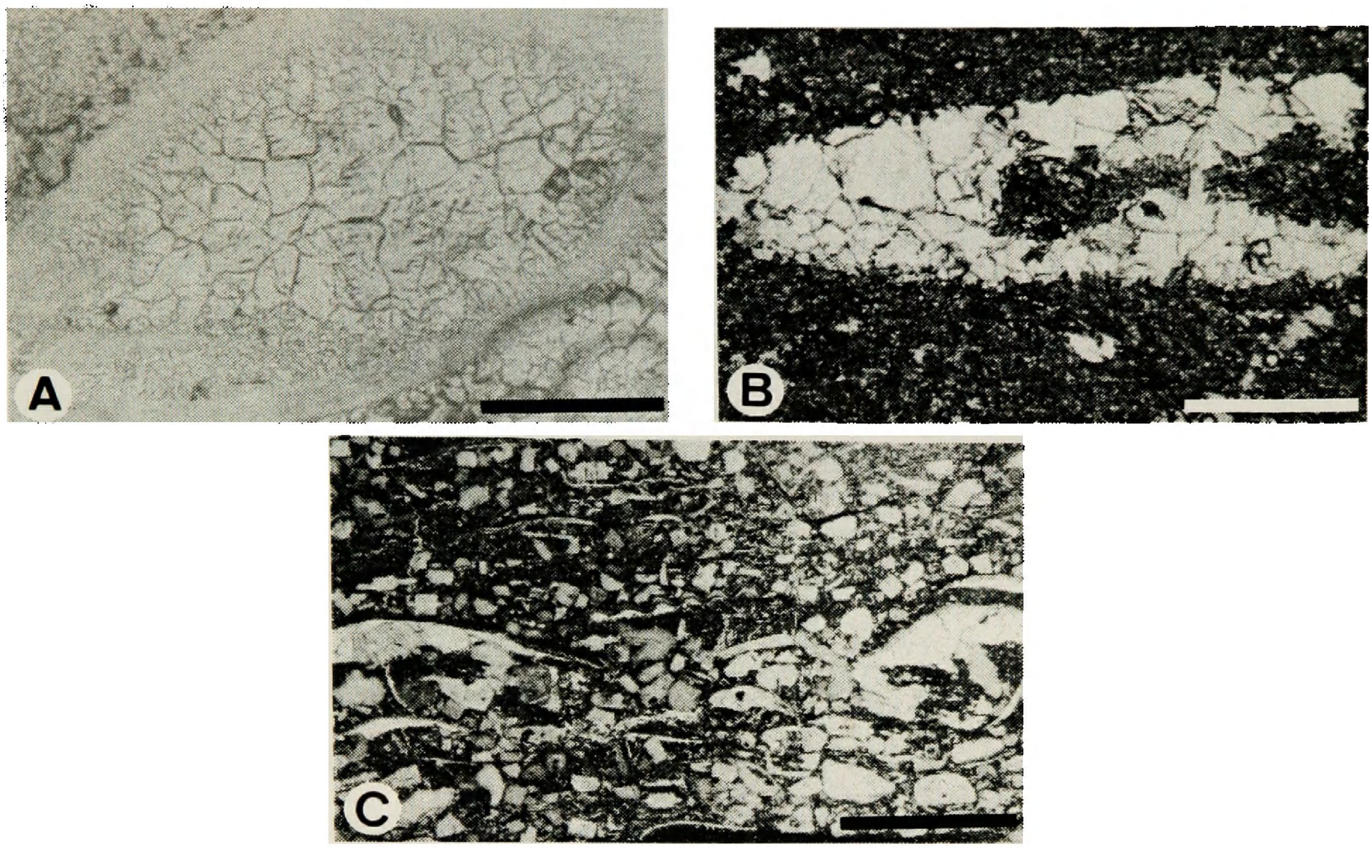


Figure 19. Common cement textures. A. Ostracod shell filled with sparry calcite cement. Acetate peel, plane light. Snowcrest Range, U.M.P. #7032. Bar=.2 mm. B. Cemented crinoid mold in calcareous, finely-crystalline dolomite. The dolomitization process in this rock dissolved the fossils (here mostly crinoids), and at one time this rock had moldic porosity. These molds were later filled-in with sparry calcite cement. Thin section, plane light. Snowcrest Range, U.M.P. #7038. Bar=.8 mm. C. Dolomitic crinoidal packstone. Blocky sparry cement occupies areas that were once void space - formed from the sheltering "umbrella" effect of the brachiopod shells. This is known as shelter cement. Thin section, plane light. Beartooth Mountains, U.M.P. #6986. Bar=10 mm. Dark area at left is from Alizarine Red-S stain.

dolomitization (fig. 19B).

In many cases, there is evidence of a time interval between the deposition of radial fibrous cement and the deposition of a coarser, blocky spar. In figure 20, it is apparent that after the radial fibrous cement grew around the ostracod(?) fragment, the sediment was subject to com-

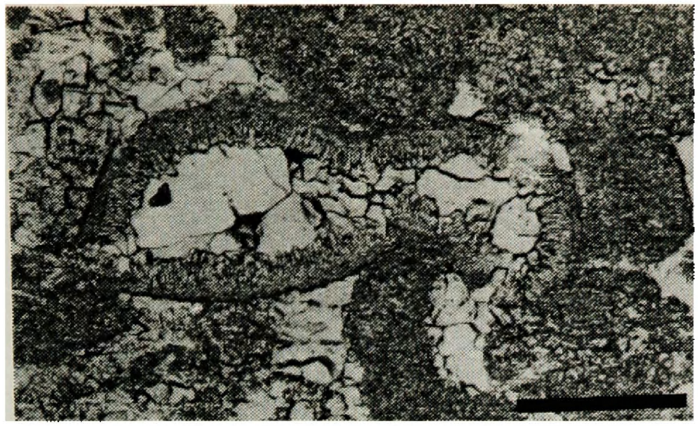


Figure 20. Shell fragment cracked by compaction. Note how the fringing cement does not extend around the cracked edges of the shell. Acetate peel, plane light. Snowcrest Range, U.M.P. #7050. Bar=.2 mm.

pressive stresses which caused the fragment to break. Later the fracture was healed by coarse, blocky spar. The evidence for this is that the fresh edges of the break do not have any radial fibrous cement on them. This same sequence of diagenetic events is also shown in figure 17 and it can be seen in other examples not illustrated. The evidence thus indicates that, most of the time, radial fibrous cement occurred early in the diagenetic history of these rocks, before compaction and before cementation by blocky, equant spar.

Rim cement on crinoid particles is cement deposited on the crinoid particle and in optical continuity with it (fig. 21). In spite of the great number of crinoid particles present in the rocks studied, rim cement is not common. Apparently, since most of the rocks studied are packstones or wackestones, the mud matrix, in contact with the

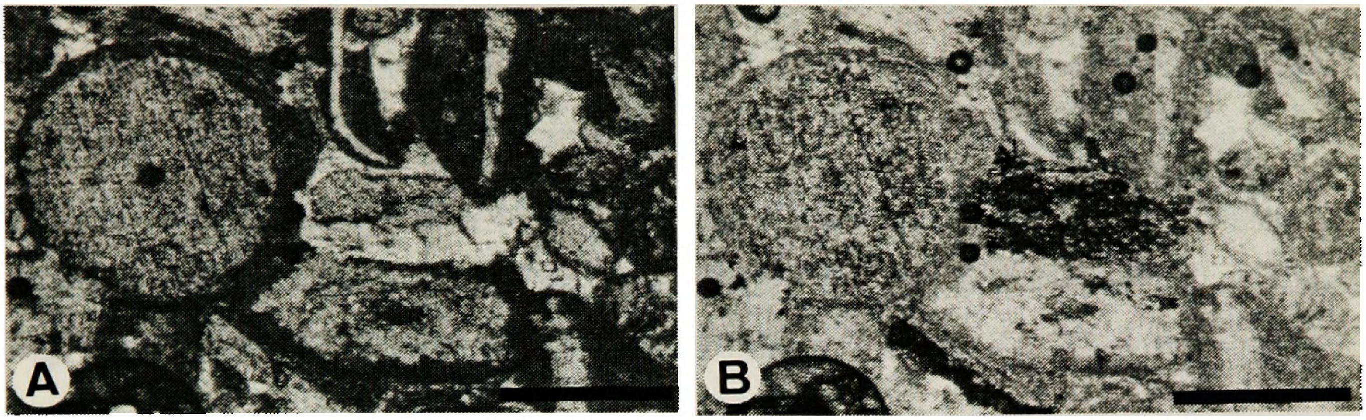


Figure 21. Rim cement on echinoderm grains. A. Plane light. The "dusty" grain in the center of the photo is an echinoderm fragment. The white is sparry calcite cement. B. Same thin section, crossed nicols. Both the echinoderm fragment and the calcite cement extinguish simultaneously. Thin section, Beartooth Mountains. U.M.P. #6982. Bar=.5 mm.

crinoid particle, prevented cement from forming. This is a common phenomenon, as noted by Lucia (1962), Evamy and Shearman (1965, 1969), and Bathurst (1975). Such cement occurs infrequently, in grainstones and in packstones that have a minimal amount of mud.

The timing of the various types of cements described above was highly variable. Radial fibrous cement occurred relatively early, before blocky sparite and before sediment compaction. Blocky sparite cements occurred somewhat later, after compaction. Sparry calcite that cemented brecciated clasts in solution breccias or karst breccias could have occurred any time from the Meramec to the early Tertiary, and there is some sparry calcite that has cemented fractures along structural cleavage planes in dolomite that is interpreted to be relatively recent.

In many carbonates, there is often more than one generation of cement. Usually, the trace geochemistry of the various generations is different, and there are various petrographic techniques useful in differentiating between generations, such as cathodoluminescence and chemical staining techniques. In the rocks studied, however, cathodoluminescence examination of thin-sections and staining techniques that might indicate multiple generations of cement (Dickson, 1965, 1966), show only one generation.

The processes and products of cementation in the rocks under study were examined as a function of diagenesis in five major environments (fig. 22). The majority of sparry

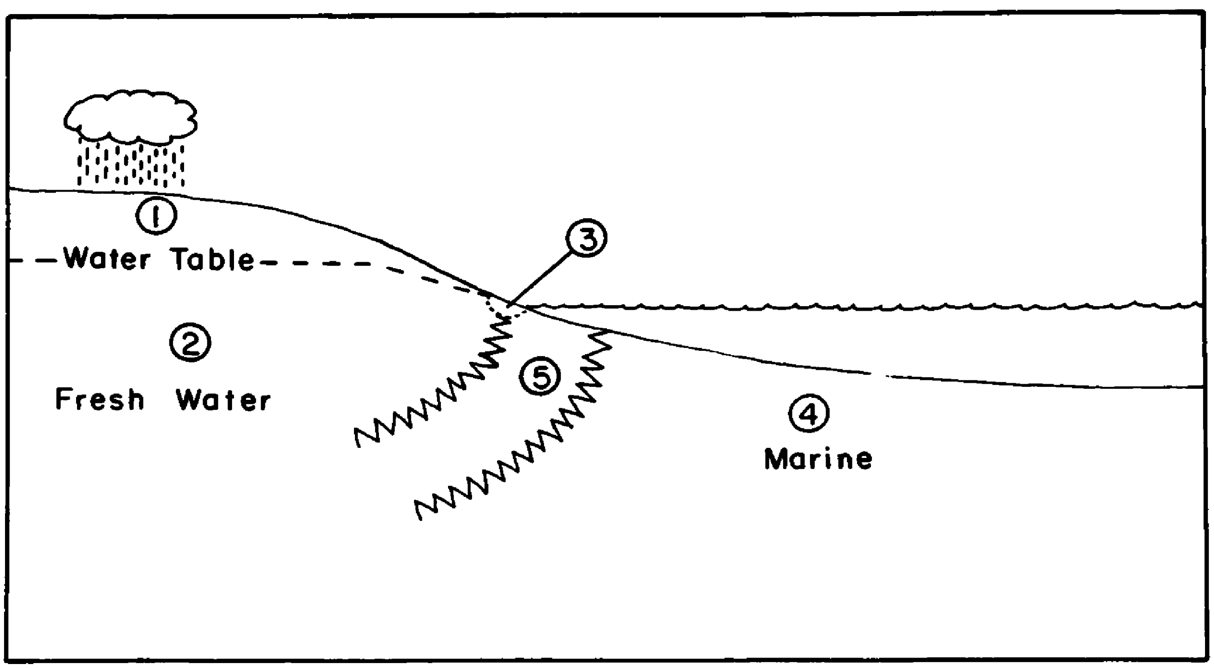


Figure 22. Major nearsurface diagenetic environments. 1. Meteoric vadose. 2. Meteoric phreatic. 3. Marine vadose. 4. Marine phreatic. 5. Mixing zone (marine phreatic and meteoric phreatic). Modified from Longman (1980).

cement textures identified indicate cementation in the meteoric phreatic zone. Evidence of fresh-water diagenesis includes: (1) rim cement on crinoids, (2) neomorphism of fossil fragments, and (3) equant calcite that coarsens toward pore centers.

Rim cement on crinoids forms very rarely, if ever, in the marine phreatic environment (Longman, 1980). In the rocks examined for this study, rim cement completely or almost completely surrounds crinoid grains wherever it occurs.

Pittman (1974) described how pore water may be undersaturated with respect to aragonite while at the same time saturated with respect to calcite. This chemistry would result in aragonite grains being dissolved, and then either being replaced or cemented by calcite. This occurs frequently in the fresh-water phreatic environment (Bathurst, 1975; Longman, 1980), and, as described previously in this paper (see fig. 10), is a common occurrence in the rocks studied.

Equant calcite cement that coarsens toward pore centers is another common feature of fresh-water phreatic diagenesis (Longman, 1980). Folk and Land (1975) described how Mg^{++} ions present in sea water poison the sideward growth of the calcite crystal and force it to grow most rapidly along its C-axis, and that only under particular circumstances will marine phreatic calcite take on an equant, blocky shape.

There is no unequivocal evidence of vadose diagenesis in the rocks studied. Meniscus cement (Dunham, 1971), as well as gravitational or pendant cement (Muller, 1971), are both diagnostic of diagenesis in the vadose environment. Their original outline may be preserved even after complete cementation (Scholle, 1978). Neither of these types of cement were observed in the rocks under study, even though a very thorough search was made for them. Land (1970), however, reported that in a tightly cemented rock, the original outline of meniscus cement (and presumably gravitational cement) may be destroyed.

The importance of submarine cementation has recently become recognized (Shinn, 1969; Milliman, 1971; Pingitore, 1971, 1976; Schroeder, 1973). Evidence for cementation in the marine phreatic zone includes fibrous cement (Longman, 1980), polygonal boundaries formed by the intersection of such cement (Shinn, 1975), microdolomite inclusions in neomorphosed radial fibrous calcite (Lohmann and Meyers, 1977), borings in cement or encrusting organisms, and pseudopellet clusters of micritic Mg-calcite (Land and Goreau, 1970). The rocks under study exhibited no evidence of marine phreatic cementation.

Because of the present poor understanding of diagenesis in the mixing zone of marine and meteoric water, it was not evaluated.

Dolomitization

The distribution and physical description of dolomite in the Mission Canyon Formation was discussed in the section on lithology (p. 21).

Much of the dolomite in the study area is believed to be of a penecontemporaneous, or early diagenetic, origin. The patterns of dolomitization and of the evolution of porosity clearly show that, initially, dolomite preferentially replaces carbonate mud rather than skeletal particles. Later on in diagenesis, depending on the original porosity of the mud and whether the carbonate available for dolomitization was derived locally or not, the skeletal particles will either dissolve or be replaced pseudomorphically (see discussion in next section). If the limestone had been cemented before dolomitization began, this pattern would not result. In those rocks that are now 100% dolomite with no hint of their original composition, this evidence does not apply.

Additionally, much of the dolomite in the rocks under study is thought to have formed in the supratidal environment, probably from water rich in Mg^{++} resulting from the precipitation of gypsum, based on the following evidence (also see Textoris, 1969, and Zenger, 1972):

A. There is a striking correlation of dolomite with evaporite-solution breccias (fig. 28). Although neither thin-section analysis nor X-ray diffraction of brecciated rocks show any suggestion of evaporite minerals, these breccias are likely the equivalent of evaporite beds in the adjacent subsurface. This has been shown in various other areas throughout Montana (see p. 10). The Centennial Range section is brecciated throughout, and this section is almost entirely dolomite. In the Snowcrest Range section, there are evaporite-solution breccias in the middle of the section, and dolomite is confined to the middle of the section. In the Beartooth Mountains section there are two evaporite-solution breccias, one at the top and one at the bottom of the section, and here, dolomite is found only at the top and bottom of the section. In the Camp Creek section, there are no evaporite-solution breccias and there is no dolomite.

B. Sections closest to the postulated ancient shoreline contain the most dolomite (fig. 1). The Centennial Range section, which is closest to shore, is 96% dolomite. The Camp Creek section, which is farthest from shore, is 100% limestone. The other two sections, at intermediate distances from shore, contain from 30 to 50% dolomite.

C. Dolomite is frequently associated with sedimentary features indicative of tidal-flat deposition. These features include algal laminations, settle-out tidal laminations, mud cracks, birdseye structures, and possible teepee structures. In addition, in the Centennial Range section, the Lodgepole Formation for at least 8.5 m below the Mission Canyon Formation is also indicative of intertidal or supratidal deposition.

Other dolomite, particularly in the sections in the Beartooth Mountains and in the Snowcrest Range, may have been dolomitized by the basinward migration of dolomitizing fluids from the shallower shelf areas, and may be diagenetic or epigenetic.

X-ray diffraction patterns of the dolomitic rocks of this study show that most of the dolomite is slightly non-stoichiometric. It ranges in composition from about 49.5 to 53 mole percent $CaCO_3$. This is well within the range of most other ancient dolomitic rocks reported in the literature (Lumsden and Chimahusky, 1980).

DOLOMITIZATION AND THE ORIGIN OF POROSITY

Porosity in the Mission Canyon Formation of southwestern Montana is almost invariably associated with dolomitic rocks. A direct relationship exists between the degree of porosity in a rock and the amount of dolomite (fig. 23). Because of this relationship, the dolomitization process is proposed to be the agent responsible for the development of porosity.

Because dolomite has a molar volume 12 to 13% greater than calcite, excess carbonate is required for the conversion of calcite to dolomite. Dolomitic rocks of this study were examined with respect to two different sources for the excess carbonate: distant source, and local source. In distant source dolomitization, carbonate ions come into the sediment along with magnesium. In local source dolomitization, mag-

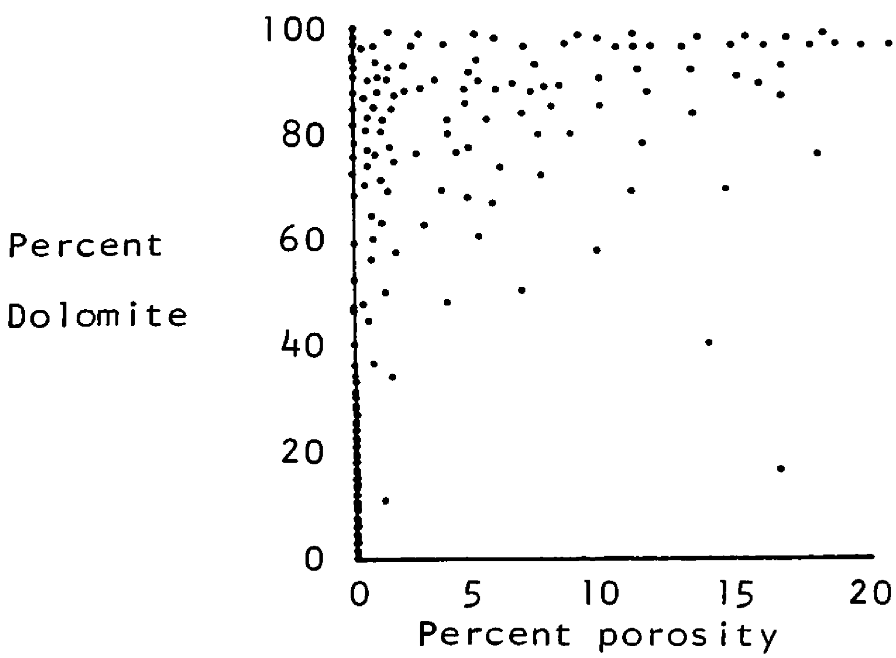


Figure 23. Percent dolomite versus amount of porosity for 159 samples.

nesium is introduced from outside, but carbonate is derived from very near the site of dolomitization (Murray, 1960; Weyl, 1960). Several observations of the rocks under study have bearing on the discussion:

A. Partially dolomitized limestones exist that were probably originally crinoidal packstones or wackestones. They have their intercrinoid area preferentially dolomitized while the crinoids are unaltered. The intercrinoid area is presumed to have been carbonate mud (fig. 24A).

B. Dolomites exist as in A above, except that the crinoids are completely dissolved away and can be identified as such only by the shape of the mold (fig. 8).

C. Dolomites exist as in A above, except that the crinoids have been dissolved, the mold filled-in by sparry calcite cement, and later replaced by several crystals of dolomite that are larger than the crystals that make up the matrix. These are identified as crinoids on the basis of the gross shape of the larger dolomite crystals (fig. 24B).

D. Dolomites exist as in A above, except that the crinoids have been completely replaced by a single dolomite crystal that is presumably in optical continuity with the original crinoid particle (fig. 24C).

The observation that the crinoids resisted dolomitization until all the mud had been dolomitized is a common one (Lucia, 1962; Murray and Lucia, 1967). This is probably a function of the permeability of the original sediment

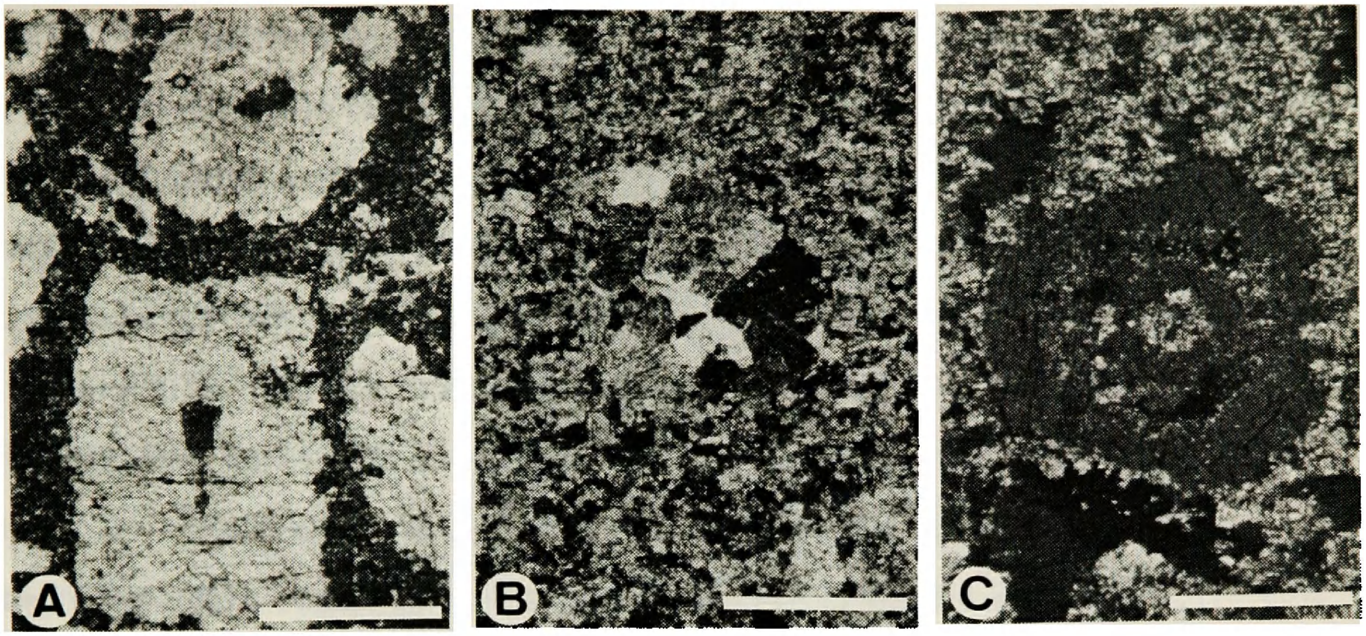


Figure 24. Crinoid grains in dolomitic rocks. A. Unaltered crinoids in a matrix of medium-grained dolomite. Original rock was probably a crinoidal packstone. Only the mud matrix has been dolomitized, including that mud in the axial canals. Thin section, plane light. Snowcrest Range, U.M.P. #7044. B. Local coarsening of dolomite crystals where they replace a cemented crinoid mold. Dolomite usually will not act as cement to fill void space. If the dolomite replaced a crinoid it would do so as a single crystal. The presence of sparry cement (white) in this structure, then, must be the remains of a cement-filled crinoid mold, and the dolomite was able to grow larger here because of the larger size of the sparry cement, compared to the micrite. Thin section, crossed nicols. Snowcrest Range, U.M.P. #7045. C. Completely dolomitized crinoid particle in a matrix of medium-crystalline dolomite. The crinoid grain here has been replaced by a single crystal of dolomite. Dark patches are pore space. Thin section, crossed nicols. Snowcrest Range, U.M.P. #7040. Bar=.8 mm in all photos.

rather than a process related to grain size, i.e., the mud was dolomitized preferentially because it was more porous and thus more likely to be exposed to dolomitizing fluids than the less porous crinoid particles.

Porosity in dolomitic rocks could have existed prior to dolomitization, or it could have been formed during or after dolomitization. Distant source dolomitization destroys porosity (Murray, 1960). Molds and other pore space that existed prior to dolomitization would probably be occluded. Dissolution of crinoids, then, must have occurred after dolomitization. With few or no exceptions, crinoid molds are only found in dolomitic rocks. It is unlikely that dissolving meteoric water would selectively leach out only those crinoids which exist in dolomite and not in limestone, especially since distant source dolomitization results in a relatively "tight" dolomite. There are certainly examples of crinoid molds that have been enlarged by solution after dolomitization, but the original formation of these molds is probably directly related to the dolomitization process.

The dolomite thus probably originated from a local source of carbonate, and the dolomitizing sequence follows that of Murray (1960) and Lucia (1962): A solution with a high Mg/Ca ratio, likely related to the precipitation of gypsum or anhydrite, permeates downward through unlithified carbonate sediments. This solution preferentially percolates through the fine-grained micritic carbonates because of their high permeability, and dolomite replaces calcite. Since dolomite has a greater molar volume than calcite, additional carbonate must be supplied to the growing dolomite from a physicochemical environment outside the area of the rhomb. In addition, more carbonate must be supplied to fill-up the void spaces of the original carbonate mud. It seems most likely that carbonate mud would be the source for the required carbonate, since crinoid grains are large single crystals of calcite, and thus are more resistant to alteration. This dissolved carbonate mud between the dolomite rhombs results in new pore space, producing intercrystalline porosity. Eventually, as the surrounding mud is utilized and dolomitization proceeds, the crinoid grains either become dissolved or replaced by dolomite. Which process occurs depends on the original porosity of the mud. If the mud is highly porous, more carbonate must be supplied, and the crinoid grains will be dissolved (fig. 8). If the mud is less porous, crinoid grains can remain whole (fig. 24A) or, with continued dolomitization, become replaced by a single crystal of dolomite (fig. 24C). Note that in the local source dolomitization scheme, porosity of the original limestone is enhanced. The resulting rock will have a porosity equal to that of the original limestone plus 12 to 13%, assuming no compaction. This also explains why unaltered crinoids, dissolved crinoids, and replaced crinoids can exist simultaneously in one rock.

Most of the replaced crinoids are replaced by a single crystal of dolomite, but some are replaced by several crystals of dolomite with different optical orientations (fig. 24B). This is not the same as the process of impingement discussed by Lucia (1962). Rather, these crystals appear to have nucleated within the solid area of the crinoid grain. This can only be explained if the area of the crinoid grain was calcite cement with varying optical orientations, i.e., the crinoid had been dissolved and the mold cemented by calcite. The dolomite then pseudomorphically replaced each individual calcite crystal within the former mold. Because of the mechanics of local source dolomitization, this must have happened after dolomitization of the mud.

If the dolomite is of a coarse grain size (500 to 1000 microns), then the ability to distinguish between former mud and grains will probably be lost, and the rocks will appear simply as a coarse-grained mosaic of dolomite. Little can be said about the precursor limestone. On the other hand, very fine-grained dolomites (less than 10 microns) also exist; some with no visible structures, and some with excellently preserved structures and fine-grained fossil fragments (fig. 25). Because these textures are so well-preserved, it is hard to envision similar fine-grained dolomite destroying other textures under similar circumstances. It is therefore proposed that very fine-grained dolomites without any visible structures or textures are that way because there were no structures or textures to preserve, i.e., they were originally fine-grained featureless mudstones.

DEPOSITIONAL MODEL

The shallowing-upward model

A geologic facies model is a general summary of a specific sedimentary environment, expressed as a block diagram, an idealized sequence of facies, or as an equation or graph. A well-developed model can be used as a framework for future observations, as a basis of comparison with other specific examples, and as a predictor in new geologic situations (Walker, 1984).

The shallowing-upward model for carbonates has been constructed from over 50 well-documented ancient and modern examples (James, 1984). Unlike terrigenous clastic sediments, carbonate sediments, particularly those deposited in shallow water, are produced directly in their environment of deposition. A healthy reef or carbonate platform can grow vertically at average rates of up to 10 mm per year (Adey, 1978). This exceeds the average rate of subsidence of the shelf or platform, which is usually between .01 and .1 mm per year (Schwab, 1976). The sediments thus accumulate in progressively shallower water until they reach sea level. This sequence is called a shallowing-upward, or regressive, sequence (Wilson, 1975). As subsidence continues and water advances over the previously deposited sediments, new carbonate sediments will begin to accumulate as before. This sequence is often repeated many times, resulting in cyclic patterns of sedimentation. These cycles are usually asymmetric in that the proximal facies of one sequence is immediately overlain by the distal facies of the next sequence, i.e., the transgressive event is not recorded in the sediments (Wilkinson, 1982). Wilson (1975) has an excellent discussion of the variations of the shallowing-upward model for rocks of various ages, and Pittman (1978) and Schlager (1981) review causes of sea level variation.

Cyclic sedimentation sequences are known from a variety of rocks of different ages and environments. Cycles have been described from the Mississippian of western Canada (Thomas and Glaister, 1960) and central Montana (Smith, 1972a), from the Silurian of New York (Shukla and Friedman, 1983), from the Pennsylvanian and Permian of Kansas (Moore, 1964), from the Permian of west Texas (Meissner, 1972), and from many other areas.

Cycles in the Mission Canyon Formation

Of the four sections examined for this study, the Snowcrest Range section and the Beartooth Mountains section exhibit poorly to moderately well-developed shallowing-upward cycles. The Centennial Range section is interpreted to have remained in shallow water throughout its depositional history, and the section along Camp Creek is thought to have remained in deep water throughout its depositional history.

These cycles were interpreted with respect to the energy required to account for the physical and biological characteristics of the rocks (Table 1). Lateral and vertical lithologic relationships were also considered in this analysis, in addition to the parameters listed in Table 1.

The section in the Centennial Range is thought to be the shallowest section examined. Evaporite-solution breccias are interpreted to be indicative of supratidal evaporite deposition, and these breccias are abundant in this section. They are often associated with birdseye structures, algal laminations, settle-out tidal laminations, and other indications of intertidal/supratidal deposition (fig. 25). This evidence, together with the abundant solution breccias, suggests extensive exposure. In addition, the regional paleogeography of the Mississippian shows this section to be nearest to shore (Rose, 1976). Cycles in this section consist mainly of tidal-flat deposits capped by evaporite-solution breccias.

The section along Camp Creek is interpreted to be the deepest water section of the four examined. According to previous interpretations (Rose, 1976), it is situated the farthest from land, and it contains no evidence of shallow water deposition; there are no evaporite-solution breccias, no algal laminations, no settle-out intertidal laminations or other intertidal features, no grainstones, birdseyes,

Sedimentary Environment	Sedimentary Structures	Fossils	Particle size
Supratidal	Solution breccias Birdseyes Some mud cracks Some laminated dolomite Lithoclasts Some mudstones Vertical burrows Fenestral fabric Algal laminations	Rare. Occasional fragments washed up by tides.	Lithoclasts to lime mud.
Intertidal	Mudcracks Settle-out laminations Churned or mottled mudstone or dolomite Burrows	Very few.	Lithoclasts to lime mud.
Shallow Subtidal	Ooids Current laminations Cross bedding	Few.	Sand.
Subtidal	Mud-rich matrix	Abundant. Crinoids and brachiopods.	Sand or larger to lime mud

Table 1. Environmental interpretation of sedimentary features of rocks in the Mission Canyon Formation of southwestern Montana.

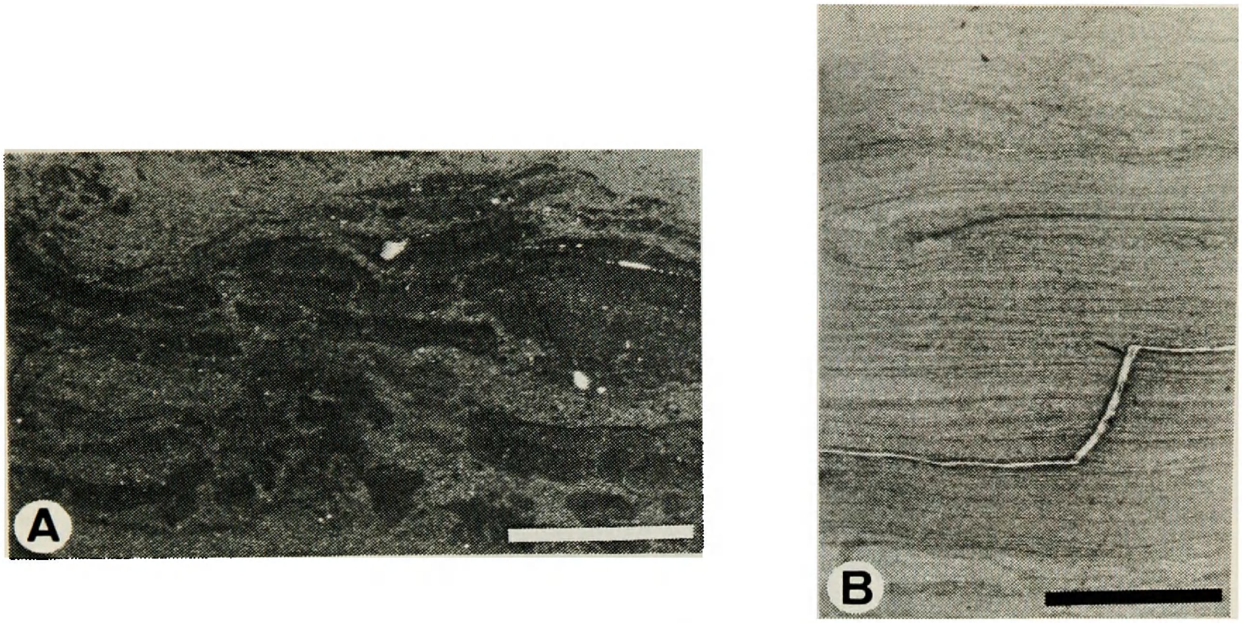


Figure 25. Well-preserved tidal features in dolomites of the Centennial Range. A. Finely-crystalline dolomite with birdseye structures and algal-laminated rip-up clasts, probably indicative of tidal-flat sedimentation. Acetate peel, plane light. Centennial Range, U.M.P. #7024. B. Finely-crystalline dolomite with 1-2 mm thick laminae of alternating finer and coarser-grained dolomite. Found immediately below rock in fig. 25A, and also probably indicative of tidal-flat sedimentation. Thin section, plane light. Centennial Range, U.M.P. #7022. Bar=10 mm in each photo.

mudcracks, etc. Because of the depth of the water, sea-level fluctuations were not reflected in the sediments, and cycles were not recorded. The rocks of this section are monotonous successions of crinoidal packstones/wackestones which have been subject to a considerable amount of recrystallization. Folding has resulted in "stretched" or otherwise deformed crinoids (fig. 14), and in the recrystallization of much of the matrix to microspar and/or pseudospar (fig. 26).

The Beartooth Mountains section and the Snowcrest

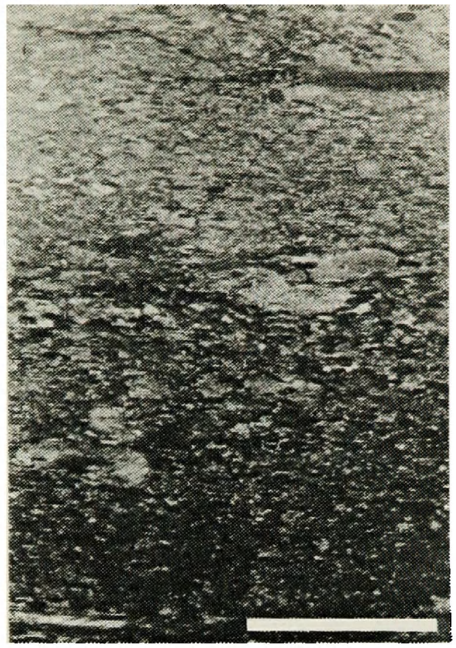


Figure 26. Tectonically deformed crinoidal packstone. Note the "stretched" crinoids, the horizontal orientation of the particles, and the abundance of microspar in the matrix. Thin section, plane light. Camp Creek, U.M.P. #7055. Bar=10 mm. Dark area in bottom half of photo is from Alizarine Red-S stain.

Range section were deposited in water intermediate in depth between these two extremes. There is evidence of shallow-water high-energy environments, shallow-water low-energy environments, and deep-water low-energy environments (plate 1). These sections exhibit poorly to moderately well preserved shallowing-upward cycles, particularly the section in the Beartooth Mountains. The cycles in these two sections are similar to the "grainstone-type" cycle described by Wilson (1975).

An idealized cycle, typical of those found in the

PLATE 1. CARBONATE ROCKS FROM THE BEARTOOTH MOUNTAINS AND THE SNOWCREST RANGE SECTIONS.

A. Faint algal-laminations in a brecciated brachiopod biosparite. Acetate peel, plane light. Snowcrest Range, U.M.P. #7031. Bar=10 mm.

B. Brachiopod crinoidal biomicrite. Dark patch in the upper left is a burrow that contains calcite after dolomite. Rock is somewhat horizontally-laminated, probably caused by currents that had enough energy to orient the particles, but not enough to remove the mud. Thin section, plane light. Snowcrest Range, U.M.P. #7035. Bar=10 mm.

C. Brachiopod shell in a crinoidal packstone. Dark grain at the upper left is a crinoid fragment. Note the spines still attached to the brachiopod shell. They, along with the shelter cement (lower right) suggest a lack of agitation of the water and quick burial of the sediment. Thin section, plane light. Beartooth Mountains, U.M.P. #6986. Bar=.8 mm.

D. Peloidal oosparite. The dark grains here are mainly ooids and skeletal fragments that have been heavily micritized. Occasionally the poorly preserved radial structure of an ooid can be seen. An early fringing cement prevented most of these particles from contacting one another when they were later compacted. Thin section, plane light. Beartooth Mountains, U.M.P. #6996. Bar=.5 mm.

E. Bottom and top sections of photo are peloidal oolitic grainstones, and the center is a mudstone. The grainstone has a somewhat gradational texture into the mudstone. The mudstone, however, is in sharp contact with the grainstone at the top, and is mudcracked. This is a good example of tidal flat sedimentation. Thin section, plane light. Beartooth Mountains, U.M.P. #6998. Bar=10 mm.

F. Lower $\frac{1}{4}$ of photo is a bioclastic packstone consisting of peloids, brachiopod fragments, and crinoid fragments, alternating in 1 mm thick laminae with mudstone. The upper part of the photo is a burrowed dismicrite with a few birdseye structures. Thin section, plane light. Snowcrest Range, U.M.P. #6989. Bar=10 mm.

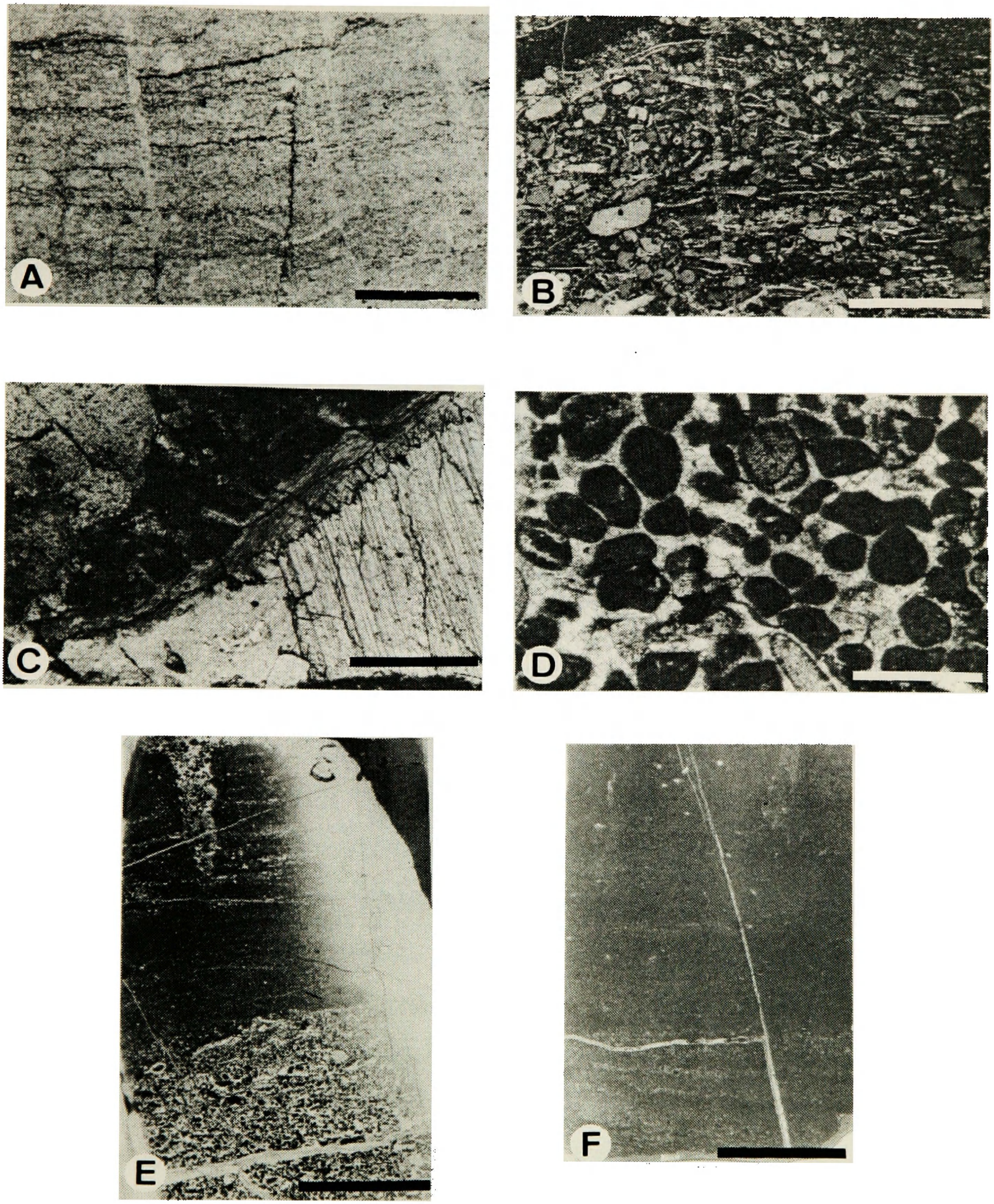


Plate 1. Carbonate rocks from the Beartooth Mountains and Snowcrest Range sections.

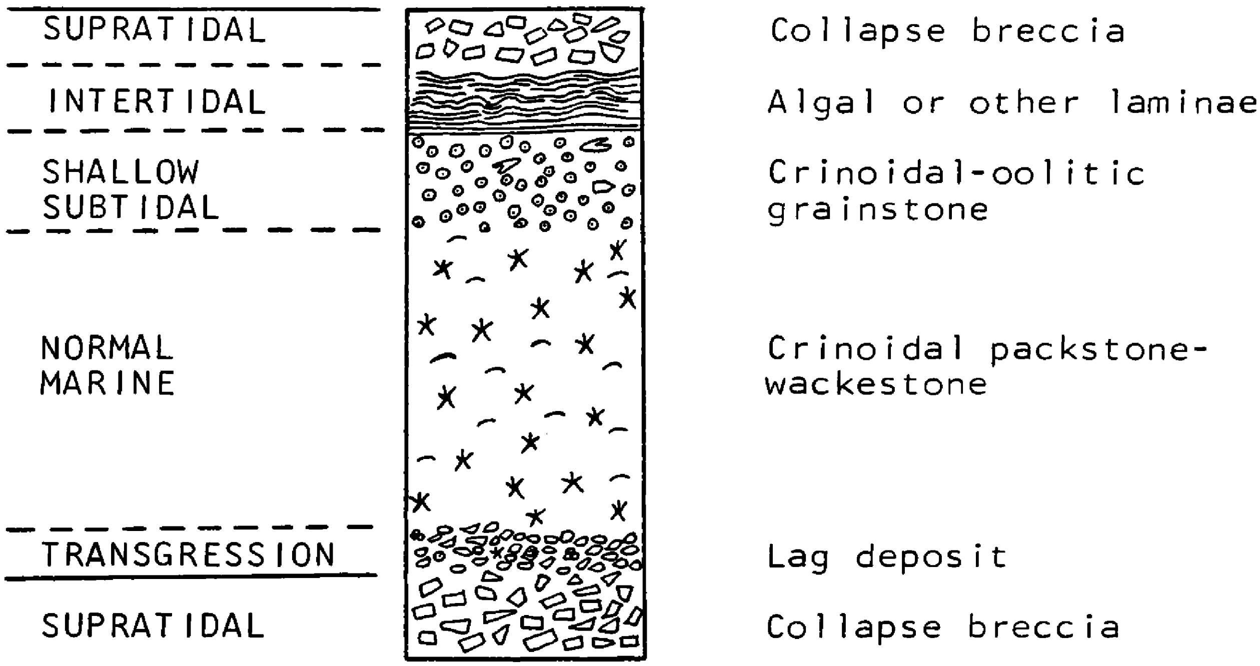


Figure 27. Typical ideal shallowing-upward cycle in the Mission Canyon Formation of southwestern Montana.

Snowcrest Range and in the Beartooth Mountains, is shown in figure 27. It would typically start with a lag deposit that indicates the initial transgression of the sea. This initial transgression probably took place quite rapidly. Rip-up clasts, intraclasts, re-worked fossil fragments, and lithoclasts are common features of this zone. These deposits are thin and often poorly preserved. Above the lag deposit a sequence of open marine or lagoon deposits exists. They are generally indicative of a fairly deep (ten to one hundred meters water depth), quiet water environment, above or below the photic zone, and below fair weather wave base, but able to be affected by occasional storms. In this environment, the most common lithology is a crinoidal packstone/wackestone, with lesser amounts of brachiopods, bryozoans, gastropods, and forams. The abundance of crinoids is indicative of open marine conditions, as crinoids are strongly stenohaline. Above this zone is the shallow subtidal zone. This zone is characterized by well-washed and sorted lime sands, commonly oolitic or pelloidal, and with no mud. These grainstones could represent either tidal channels which exist on the tidal flat, or sandy shoals existing seaward of the beach. In either case, these grainstones could be found stratigraphically above or below the intertidal sediments (James, 1984). The intertidal zone is characterized by algal laminations, intertidal "settle-out" laminations (Hardie and Ginsburg, 1977), mud cracks, birdseyes, graded storm layers, possible teepee structures and stromatactis, and dried-out chips of lime mud. There is also a general lack of fossils, except for a few that may have been washed up by the tides. The supratidal environment is identified mainly on the basis of the evaporite-solution breccias, as they are believed to result from the collapse of sediments superjacent to an evaporite layer after the evaporites dissolved. This breccia marks the top of the cycle.

Only rarely are all of these elements present in any one cycle. The initial lag deposit, particularly, is often not well preserved, and supratidal features, other than collapse breccias, are sometimes missing.

Such cycles are best developed in the Beartooth Mountains, where at least four and possibly five or six cycles can be seen (Appendix C). Cycles here average 42 m thick, and typically shallow upward to an oolitic grainstone or to an oolitic grainstone and an evaporite solution breccia. Where a breccia is not present, the seas apparently did not shallow enough for a supratidal sabkha to develop. The top 15 m of this section appears to have remained in very shallow water (intertidal to shallow subtidal).

These cycles are also seen in the Snowcrest Range, although they are not as numerous nor as evident as in the Beartooth Mountains. Here, there is at least one cycle, and possibly three or four. Cycles in this section average 47 m thick, and they shallow upward to an evaporite-solution breccia with no grainstone, or to a crinoidal grainstone above an evaporite-solution breccia. The more abundant evaporites in this section indicates that it may have been closer to shore than the Beartooth Mountains section. As in the Beartooth Mountains section, the top 21 m of this section remained in very shallow water.

In the Centennial Range, cycles average 30 m in thickness and consist of tidal-flat deposits capped by evaporite-solution breccias. An oolitic grainstone exists at the bottom of the section that may be a beach sand marking the shoreline during the initial transgression of the sea.

Depositional model

The inferred depositional model is shown in figure 28. With section 2 (Centennial Range section) remaining intertidal/supratidal most of the time, and section 4 (Camp Creek section) remaining in deep water, section 1 (Beartooth Mountains section) was in an area of offshore shoaling, and section 3 (Snowcrest Range section) was shoreward of the shoal.

For the sections studied, the development of a single cycle through time could thus be presented as follows: The initial transgression of the sea drapes a thin lag deposit across the land surface until the sea stabilizes at the stage depicted in figure 28A. Here the maximum wave energy impinges on the shore, creating a well-washed oolitic carbonate beach sand and a tidal flat. This beach sand is recognized at section 2 only at the base of the section. At this stage the model may be termed a "carbonate ramp" (Ahr, 1973; original reference not seen, see Wilson, 1975). As the carbonate platform begins to build up, a shoal develops offshore, and the carbonate ramp rapidly evolves into a carbonate shelf (fig. 28B). This shoal consisted of well-washed oolitic sands, probably most well developed on its seaward side, with a well-developed tidal

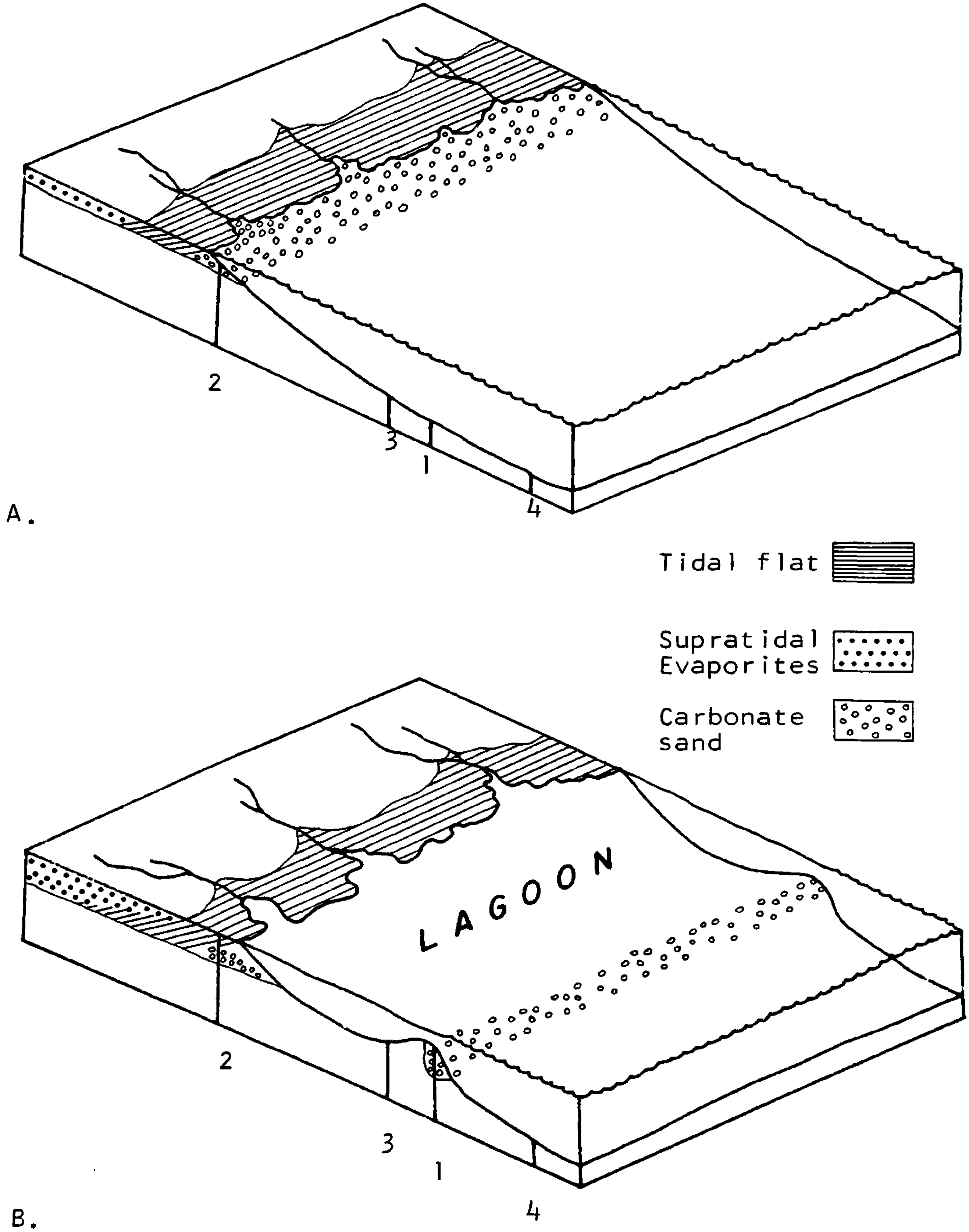
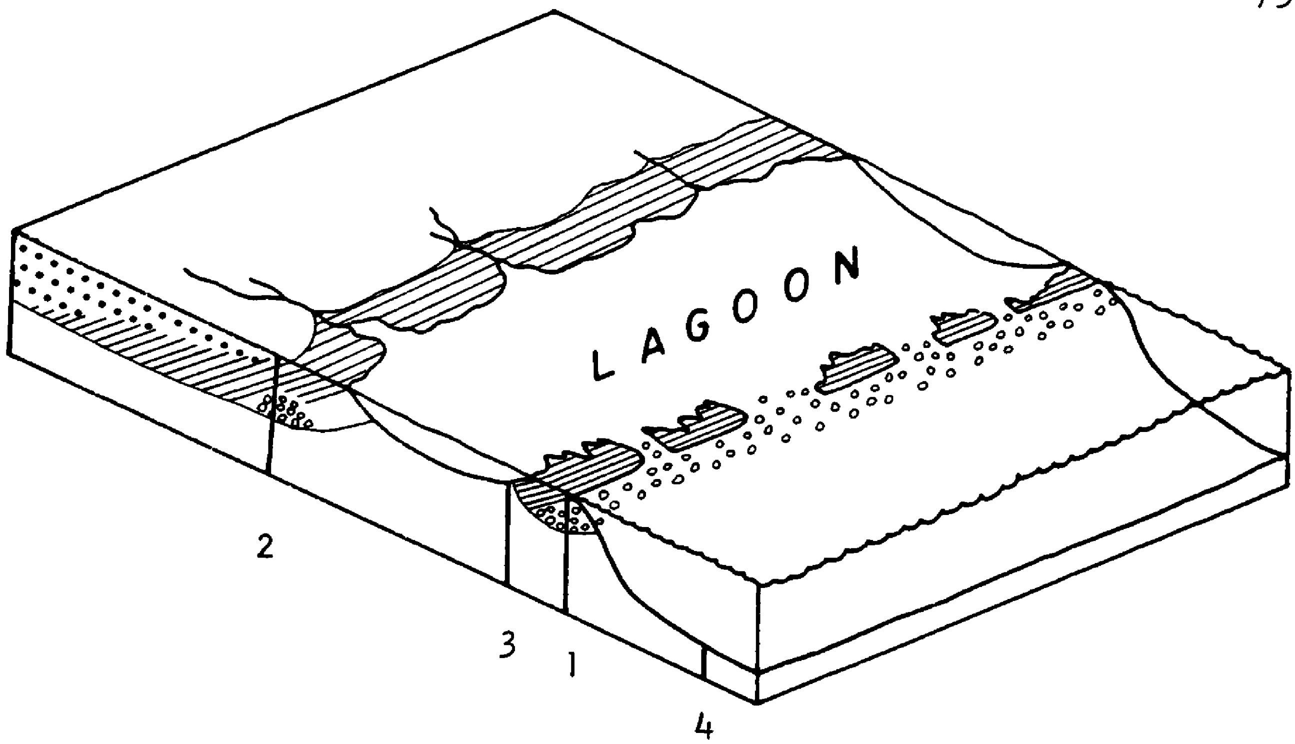
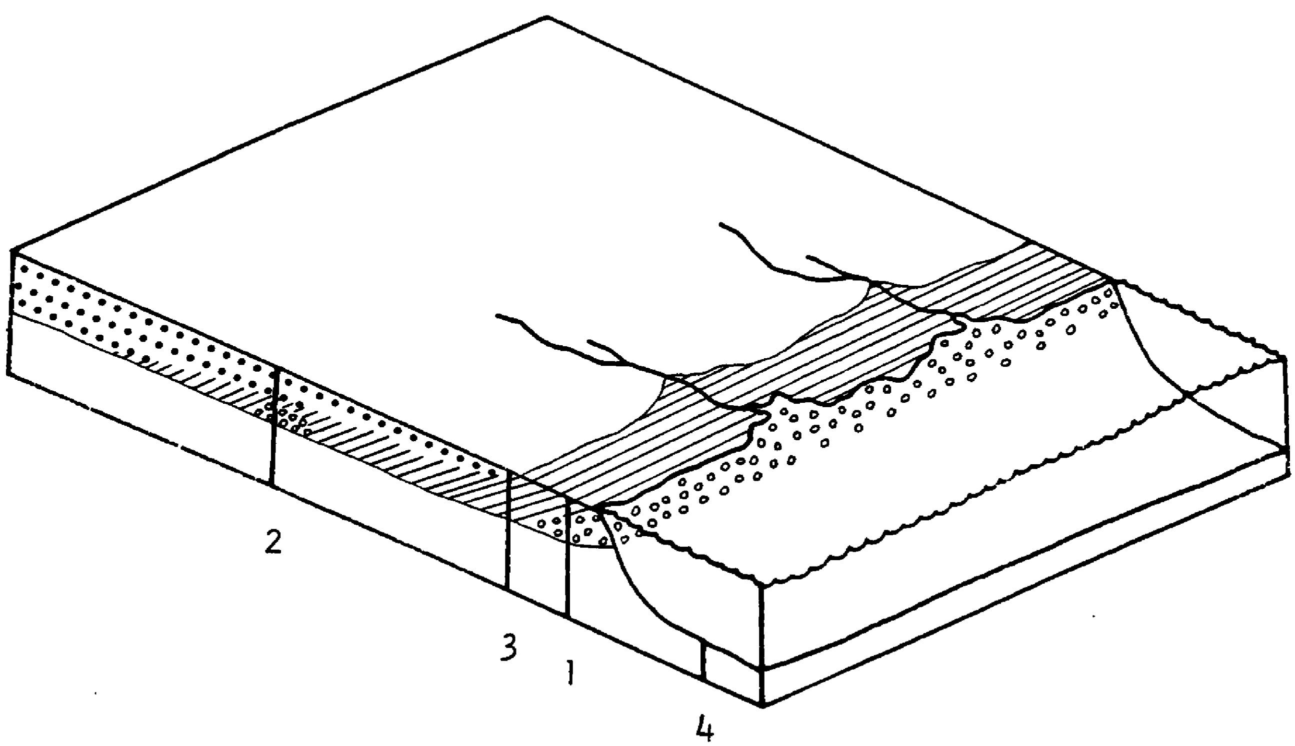


Figure 28. Depositional Model.



C.



D.

Figure 28. Depositional Model (continued).

flat on its landward side. It received most of the wave energy of the open ocean, so that for the rest of this cycle, the water shoreward of the shoal remains calm. At this time, section 4 is in deep water, section 1 and section 3 are in water somewhat shallower than at section 4, and section 2 is intertidal. With maximum development of the shoal, a well-developed oolitic sand bar will exist offshore (fig. 28C). The lagoon consists of very quiet, normal marine water which is continuously becoming shallower. At this point, section 4 is still in deep water, and section 2 is in a supratidal evaporite setting. Section 1 is in the area of shoaling with either grainstones or tidal flat sediments being deposited, depending on the exact position of the shoal. Section 3 is still shoreward of the shoal in quiet, relatively deep water, but, depending on the exact location of the shoal, would occasionally be in an area of tidal sediment or grainstone deposition.

With continued progradation of the carbonates, a supratidal sabkha is developed (fig. 28D). Section 4 is in deep water, sections 2 and 3 are supratidal, and section 1 is right on the boundary and may be supratidal, intertidal, or shallow subtidal, depending on exact conditions. This constitutes one complete cycle, and the next transgression starts the process all over again.

Diagenetic processes should differ depending on where a section is located with respect to this model. Penecontemporaneous dolomitization might be expected in shallow water sections, and in the sections studied, it does occur. Sections 1 and 3 might be expected to show the effects of fresh-water diagenesis and of penecontemporaneous dolomitization, particularly at and below the shallowest portion of each cycle. Fresh-water phreatic diagenesis is common throughout sections 1 and 3, and there is also a strong correlation between dolomitization and the sections of rocks that are found at and below zones of evaporite-solution brecciation (fig. 29). Unfortunately, section 4 has been so recrystallized that little can be said of its depositional or diagenetic history.

Correlation of sections

Correlation of these sections based on the observed cycles can be risky. Sections 1 and 3, although interpreted to exist in environments that are close to or adjacent to one another, are actually separated by a distance of 180 km. Sections 2 and 3 are close enough together, but only the evaporite-solution breccias seem to be able to be used in correlation. Section 4, the deeper water section, could not be correlated with any other section. A tentative correlation was made between sections 1 and 3, however. This may be a valid correlation because these two sections have the greatest amount of lithologic similarity, they are both from a region where cycles are developed, and they can

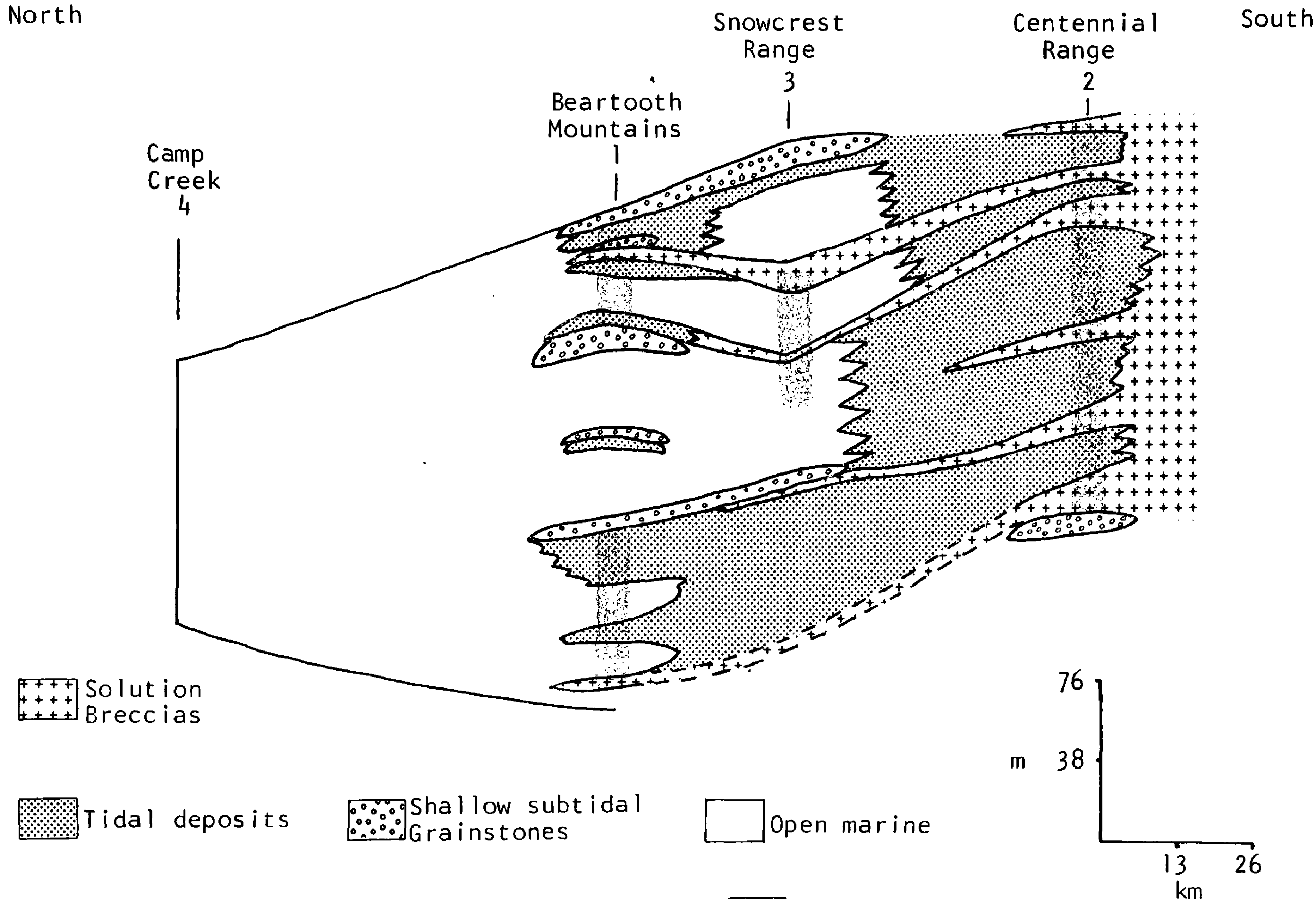


Figure 29. Correlations between sections.

also be correlated on the basis of lithology in addition to the cycles. This is in fact exactly how these two sections were correlated in the first place (see fig. 29), and that the cycles also correlate in this scheme shows that they test well (also see Hall, 1952). These cycles are, however, of little value in regional correlation. The completely marine nature of these rocks and the lack of key marker beds limits their usefulness in this regard. They lack the lithologic distinctiveness and diversity of, say, the Pennsylvanian cyclothems of the midcontinent region, with their marine and continental sediments.

Even though the carbonate buildup proceeds more rapidly than basin subsidence, the initial transgression of the sea, at the beginning of each cycle, is probably more rapid than either. Thus, points established by the boundary between the shallowest part of one cycle and the deepest part of the next cycle (the lag deposit) are probably the best for correlation. Evaporite-solution breccias may also make a good correlation line between sections. They occur at the shallowest part of a cycle, and as they are preserved more often in the sedimentary record than the lag deposit, may prove more useful in this respect.

Cycles in the Lodgepole Formation

Smith (1972a, 1972b, 1977) described cycles from the Lodgepole Formation of the Madison Group of central Montana which are somewhat different than those in the Mission Canyon. The Lodgepole Formation of his study is situated on the unstable shelf of Montana. It is characterized by a shallow-water environment represented by the Cottonwood Canyon Member, a deep water environment represented by the Paine Member, and another shallow water environment represented by the Woodhurst Member. Cycles noted by Smith (1972a, 1977) were of two distinct types: a mudstone or wackestone unit capped by an oolitic grainstone that represents the deep water environment of the Paine Member, and a pellet grainstone unit capped by oolitic grainstones that represents the shallower water environment of the Woodhurst Member. Smith (1972a) interpreted the Paine Member to be a deep water, open shelf deposit. Cycles in the Mission Canyon Formation do not contain any elements interpreted to be as deep as the ones reported in the Paine Member. The cycles in the Woodhurst Member are more similar to the ones in the Mission Canyon Formation, except that they are of a much higher average energy, as exemplified by the abundance of grainstones and the lack of mud, and they do not contain any tidal elements. The evaporites that cap some of the cycles in the Mission Canyon Formation are nonexistant in the Lodgepole Formation. This is as expected, as the Lodgepole Formation was deposited by a transgressing sea rather than by a regressing sea as was the Mission Canyon Formation.

SUMMARY AND CONCLUSIONS

1. The Mission Canyon Formation in southwestern Montana is a thin-bedded to massive sequence of limestone and dolomite, with very minor amounts of terrigenous clastics.

2. The formation has a relatively small degree of lithologic diversity. Seven major lithotypes are recognized, with crinoidal packstones, oolitic or crinoidal grainstones, and dolomite being the most abundant rock types.

3. Fossil diversity is relatively low. Echinoderms are, by far, the most abundant fossil. Brachiopods, bryozoans, forams, and ostracods occur infrequently, and gastropods and corals are even more rare.

4. The rocks studied were generally deposited in a shallow sea of normal salinity.

5. A variety of diagenetic modifications affected these rocks, including silicification, cementation, compaction, and dolomitization. Additionally, some of the rocks studied have been extensively recrystallized.

6. Cement fabrics indicate that cementation occurred in the meteoric phreatic zone.

7. Much of the dolomite is of a penecontemporaneous supratidal origin, and was effected through a local source of carbonate.

8. The process of dolomitization resulted in a considerable amount of porosity. Porosity exists mainly as molds and

solution-enlarged molds of crinoids, and as intercrystalline pores between dolomite rhombs.

9. The four sections measured delineate a carbonate platform with an offshore oolitic bar.

10. Several shallowing-upward cycles are noted in these sections. Each one generally consists of open marine deposits shallowing to an oolitic or crinoidal grainstone, a tidal zone, or a supratidal zone.

11. Evaporites were formed in a supratidal sabkha. A sabkha probably existed at various locations throughout the period of deposition of the Mission Canyon Formation.

12. Evaporite-solution breccias were formed when near-surface waters dissolved the evaporites and the overlying sediments collapsed.

13. At the end of the deposition of the Mission Canyon Formation, the seas withdrew to the west, and a karst topography was developed at the top of the section. Later deposition of the overlying sediments collapsed the cavern roofs of the karst topography, resulting in karst breccias.

14. The extremely varying thickness of the Mission Canyon Formation is primarily a consequence of the variable amount of erosion that took place at the top of the section during this exposure.

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APPENDICES

Appendix A - Stratigraphic sections.

The locations of the stratigraphic sections measured for this study that appear on the following pages are as follows:

Section 1. Baker Mountain, Beartooth Range. Approximately 30 miles southwest of Big Timber, Montana, on Route 298. Section is west of road along east side of Baker Mountain in $W\frac{1}{2}NW\frac{1}{4}$ sec. 35 and $NE\frac{1}{4}NE\frac{1}{4}$ sec. 34, T3S, R12E.

Section 2. Odell Creek Canyon, Centennial Range. Approximately three miles southeast of Lakeview, Montana, on the east side of Odell Creek in $SW\frac{1}{4}$ sec. 31, T14S, R1W.

Section 3. Snowcrest Range. Approximately 45 miles southeast of Dillon, Montana, along West Fork Blacktail Deer Creek. Section is east of road in $SE\frac{1}{4}SE\frac{1}{4}$ sec. 15, and $SW\frac{1}{4}SW\frac{1}{4}$ sec. 14, T12S, R6W.

Section 4. Camp Creek. Approximately two miles east of Melrose, Montana, along the south side of Camp Creek in $N\frac{1}{2}N\frac{1}{2}$ sec. 30, T2S, R8W.

EXPLANATION FOR STRATIGRAPHIC COLUMNS

Limestone

Dolomite

Shale

Breccia

Chert

Scale is in meters

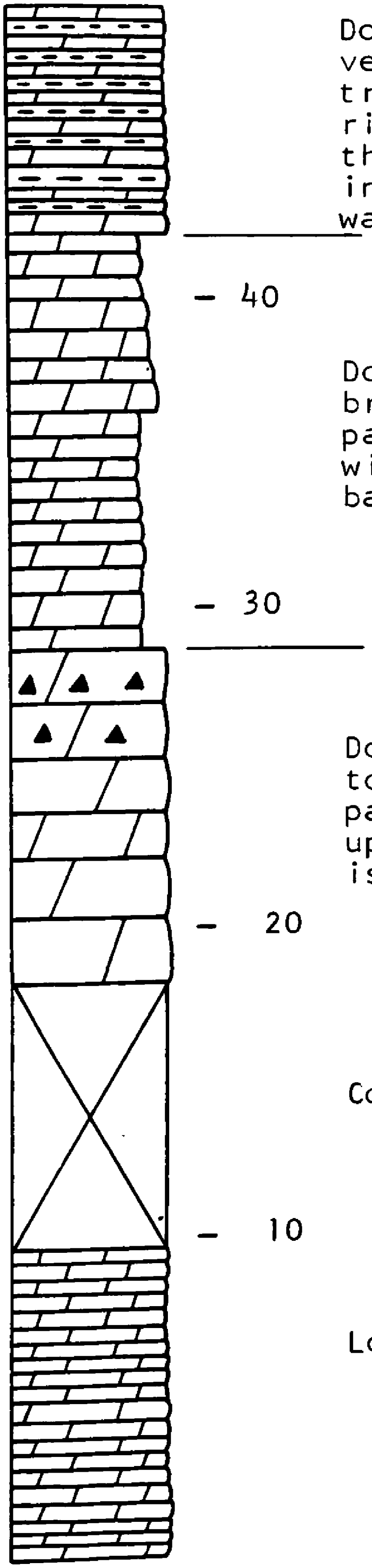
Crinoidal limestone or dolomite

Oolitic limestone

Sandstone

Covered interval

Bedding thickness:
very thin: less than 8 mm
thin: 8 mm - .3 m
medium: .3 m - .6 m
thick: .6 m - 1.5 m
very thick: 1.5 m - 3 m
massive: greater than 3 m



Dolomite; finely crystalline. Light gray, very thin bedded, with 12% scattered detrital quartz silt; cross-bedded and with ripple marks. Alternates with 3 to 5 cm thick layers of light gray very thinly laminated shale. Quartz content decreases toward the top. Mottled in places throughout.

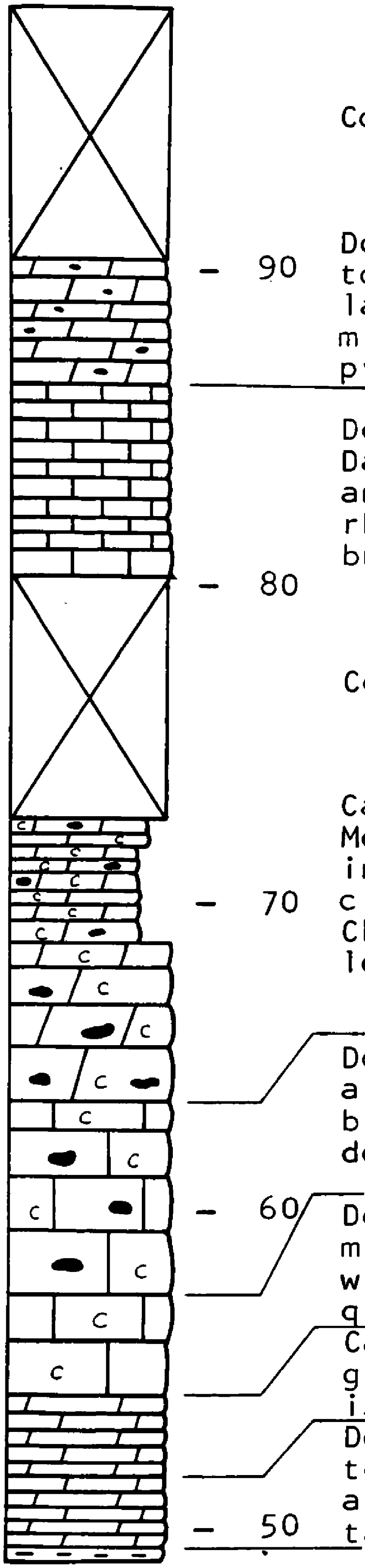
Dolomite, finely crystalline. Light grayish-brown, thin to medium bedded. Occasional patches of coarser dolomite, and 1% pyrite with associated limonite staining. Bioturbated in places.

Dolomite; fine to medium crystalline. Medium to dark gray, massive. Mottled in places with patches of coarser dolomite and sparite, with up to 10% intercrystalline porosity. Top 3m is brecciated.

Covered.

Lodgepole Formation.

Section 1 - Beartooth Mountains



Covered.

Dolomite; finely crystalline. Dark gray, thin to medium bedded. Rock is horizontally wavy-laminated with darker brown patches of dolomite. Trace of intercrystalline porosity, pyrite, and chert.

Dolomitic limestone; micrite (mudstone). Dark gray, thin to medium bedded. 70% micrite and/or microspar with scattered dolomite rhombohedra. Top 3 m are laminated and burrowed.

Covered.

Calcareous dolomite; finely crystalline. Medium gray and massive at the bottom; becoming thin-bedded the top 3 m. Contains 30% crinoids and sparite in a dolomite matrix. Cherty in places, and vaguely burrowed, mottled, and laminated at the top.

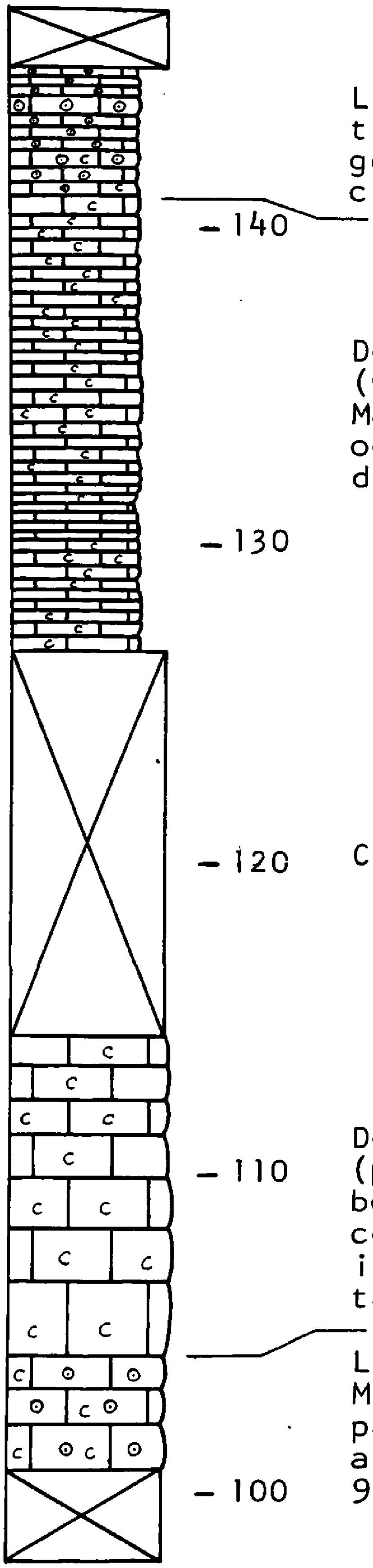
Dolomitic limestone; dolomitic cherty crinoidal biomicrite (packstone). 30% crinoids and bryozoans in a matrix of microspar with 10% dolomite. Cherty in places.

Dolomitic limestone; dolomitic crinoidal biomicrite (packstone). Medium gray, massive, with a trace of detrital and authigenic quartz, and 40% dolomite.

Calcareous dolomite; finely crystalline. Dark gray, thin to medium bedded, with 20% dolomitized crinoid particles and 15% sparry calcite.

Dolomite; finely crystalline. Dark gray, thin to medium bedded. Trace of detrital quartz and authigenic pyrite, with 1-5% intercrystalline porosity.

Section 1 - Beartooth Mountains



Limestone; Oosparite (grainstone). Dark gray, thin bedded. Ooids have both radial and tangential structure, and many have a nucleus of crinoid fragments. Trace of detrital quartz.

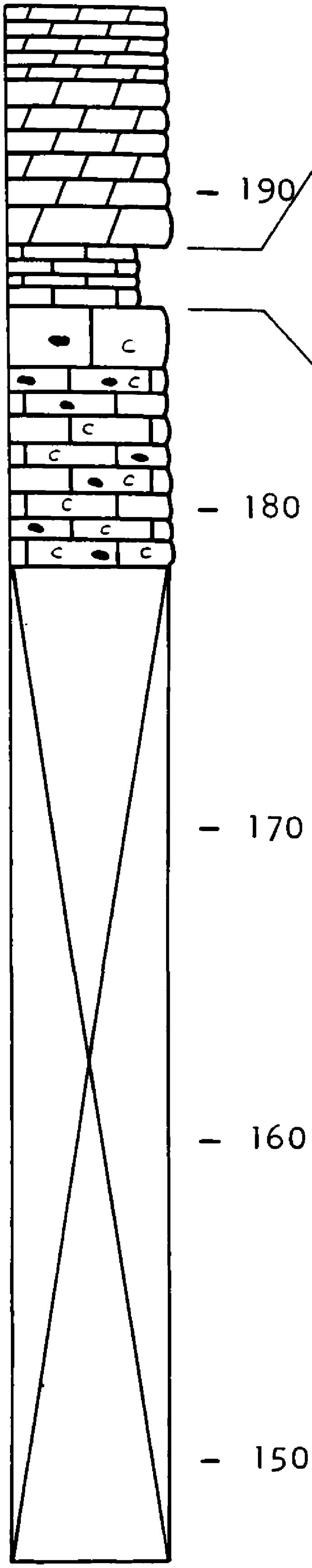
Dolomitic limestone; crinoidal biomicrite (packstone). Medium gray, thin bedded. Matrix is mostly microspar, with a few percent ooids, cortoids, and peloids. Becomes less dolomitic toward the top.

Covered.

Dolomitic limestone; crinoidal biomicrite (packstone). Dark gray and massive at the bottom; medium bedded the top 5 m. Dolomite content decreases toward the top. Lower part is horizontally laminated and burrowed. Contains about 50% crinoids.

Limestone; crinoidal oosparite (grainstone). Medium gray, massive. Stylolitic with a compacted fabric, many microstylolitic contacts, and stylolitic porosity. Consists of over 90% particles, mostly crinoids.

Section 1 - Beartooth Mountains



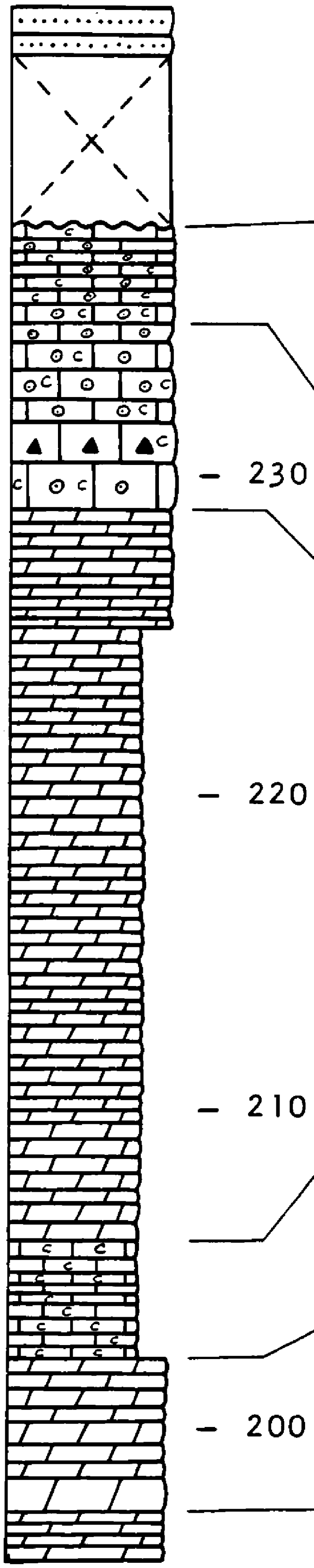
Dolomite; finely crystalline. Medium to light gray, thick bedded at the bottom, thin bedded the top 3 m. Contains rip-up clasts and 1 to 3% moldic porosity.

Dolomitic limestone; biomicrite (packstone/wackestone). Medium to light gray, thin bedded. Mudstone alternates with crinoidal brachiopod packstone in 1 mm beds, and grades into mudstone at the top.

Limestone; biosparite (grainstone). Dark gray, thin to medium bedded, becoming massive the top 2 m. Somewhat compacted fabric with trace of authigenic pyrite and chert, and 20% porosity as vugs and solution-enlarged molds.

Covered.

Section 1 - Beartooth Mountains



Amsden Formation.

Covered. Exact stratigraphic distance unknown.

Unconformity.
Limestone; oolitic crinoidal biosparite (grainstone). Medium gray, thin to medium bedded. The lower 6 m alternates with mudstone in 30 mm sets; grading into fine laminae the top 5 m.

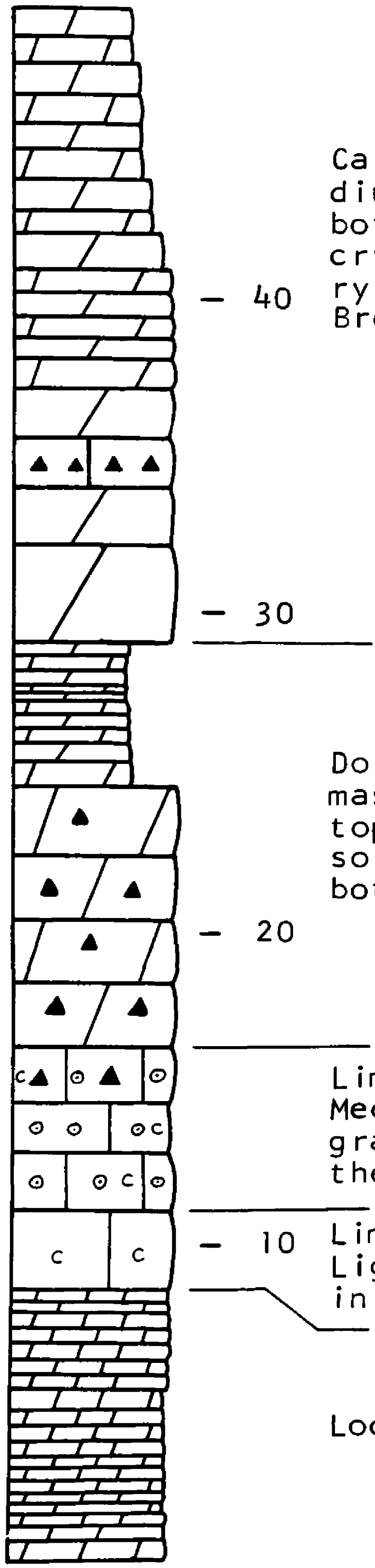
Limestone; oolitic crinoidal biosparite (grainstone). Light gray, thick bedded, with 5% scattered dolomite. Particles are well rounded and sorted, and are somewhat compacted. Top 4 m become medium bedded.

Dolomite; fine to medium crystalline. Medium gray, thin to medium bedded. Middle section is mottled from local coarsening of dolomite, and has birdseyes and intercrystalline porosity. Upper 2 m are calcareous and have a more finely crystalline texture.

Limestone; crinoidal pelmicrite (packstone). Medium to light gray, thin bedded. Matrix consists of compacted pellets with 50% crinoids and 5% dolomite.

Calcareous dolomite; finely crystalline. Medium to light gray, thin to medium bedded. 3% vuggy porosity increases to 12% the upper 2 m. Horizontally laminated at the top.

Section 1 - Beartooth Mountains



Calcareous dolomite; medium crystalline. Medium gray, massive to thick-bedded at the bottom, medium bedded at the top. 10% intercrystalline and vuggy porosity, and 10% sparry calcite replacing or cementing fossils. Brecciated at 34-35 m.

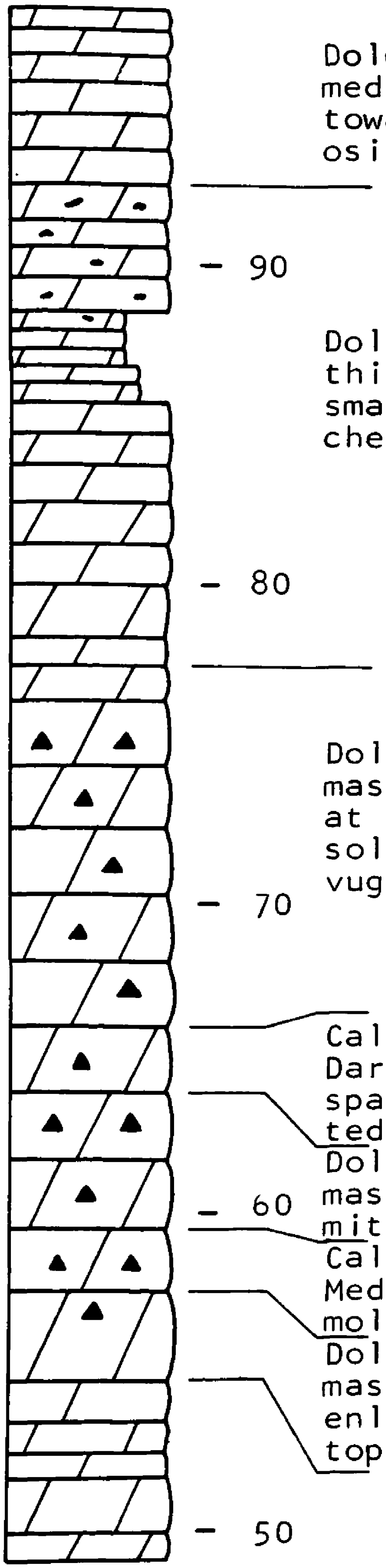
Dolomite; finely crystalline. Medium gray, massive at the bottom and medium-bedded the top 4 m. 10% porosity as vugs, molds, and solution-enlarged molds. Erecciated at the bottom.

Limestone; crinoidal oosparite (grainstone). Medium gray, massive. 75% ooids and crinoid grains. Slightly compacted. Brecciated at the top.

Limestone; crinoidal biomicrite (packstone). Light gray, massive. Consists of 40% crinoids in a micrite matrix. 10% vuggy porosity.

Lodgepole Formation.

Section 2 - Centennial Range



Dolomite; finely crystalline. Medium gray, medium bedded at the bottom, becoming massive toward the top. 15% intercrystalline porosity and as solution-enlarged molds.

Dolomite; finely crystalline. Medium gray, thin to very thick-bedded. 5% porosity as small vugs and crinoid molds. Contains small chert nodules toward the top.

Dolomite; finely crystalline. Medium gray, massive at the bottom, becoming thick-bedded at the top. 15% porosity as molds of crinoids, solution-enlarged molds of crinoids(?), and vugs. Brecciated throughout.

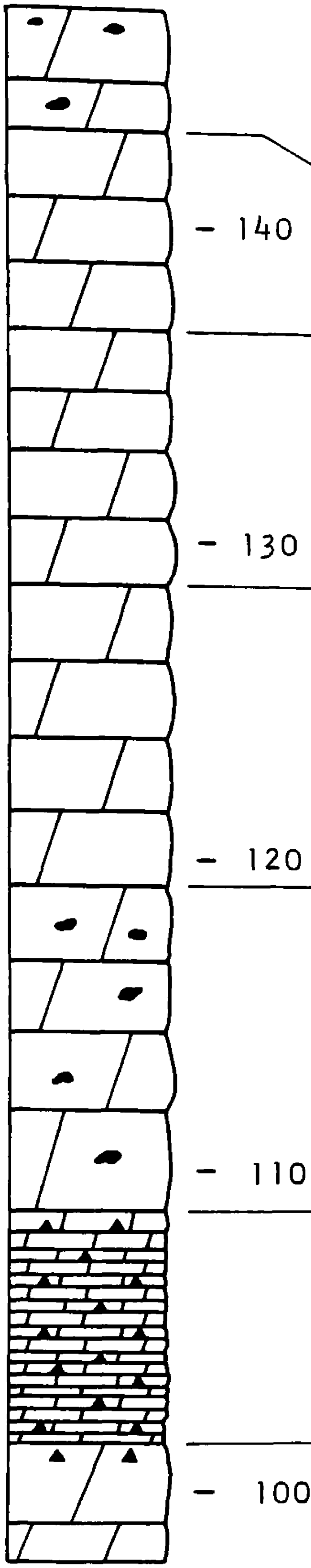
Calcareous dolomite; very finely crystalline. Dark gray, massive. 7% moldic porosity and 8% sparite as cemented molds and vugs. Brecciated throughout.

Dolomite; medium crystalline. Medium gray, massive. 10% vuggy porosity and 15% dolomitized crinoid particles. Brecciated.

Calcareous dolomite; finely crystalline. Medium gray, massive. 15% sparite cementing molds and vugs. Brecciated throughout.

Dolomite; finely crystalline. Dark gray, massive. 25% porosity as molds, solution-enlarged molds, and vugs. Brecciated at the top.

Section 2 - Centennial Range



Dolomite; medium crystalline. Light gray, massive. 15% vuggy porosity. Chert nodules throughout.

Dolomite; finely crystalline. Medium gray, massive. 3% vuggy and moldic porosity.

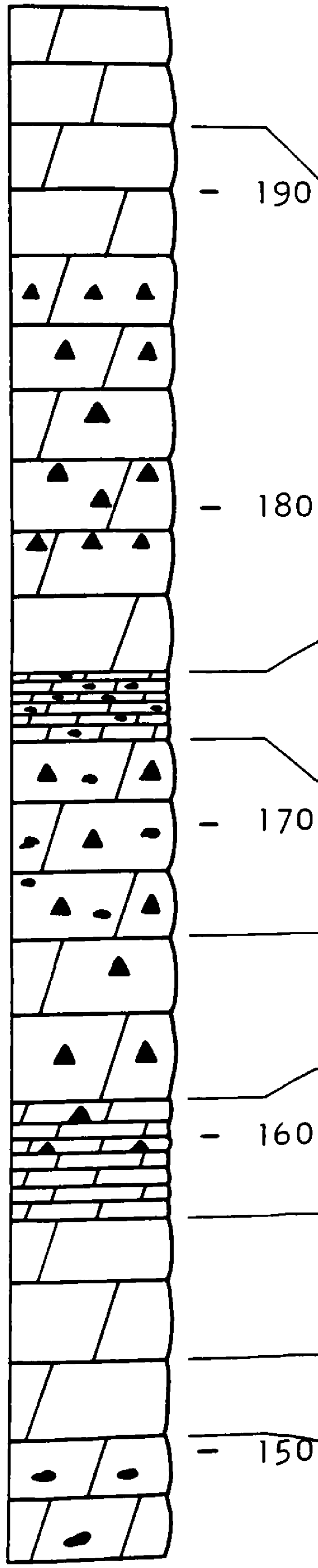
Dolomite; medium to coarsely crystalline. Light gray, massive.

Dolomite; finely crystalline. Medium gray, massive. 2% vuggy porosity.

Dolomite; finely crystalline. Medium gray, massive. 20% porosity as vugs, molds, and solution-enlarged molds of crinoids. Chert nodules throughout.

Calcareous dolomite; medium crystalline. Medium gray, thin-bedded. 10% intercrystalline porosity and as vugs of unknown origin. Brecciated throughout.

Section 2 - Centennial Range



Dolomite; medium crystalline. Light gray, massive. 10% intercrystalline porosity and 20% porosity as vugs and solution-enlarged molds of crinoids(?).

Dolomite; finely crystalline. Light gray, massive. 7% vuggy and moldic porosity. Algal laminated with birdseye structures and rip-up clasts at 180 m. Brecciated throughout.

Dolomite; finely crystalline. Light gray, thin-bedded. 1 to 2 mm thick laminations of alternating coarse (35 micron) and fine (20 micron) dolomite subhedra. Cherty throughout.

Dolomite; medium crystalline. Light gray, massive. Cherty and brecciated with up to 20% fracture porosity. Brecciated throughout.

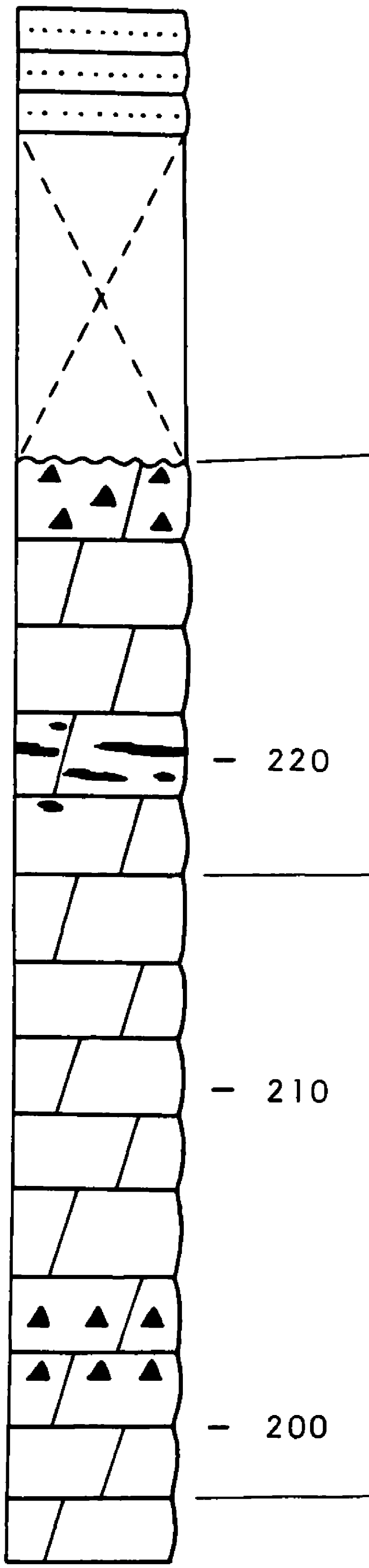
Dolomite; coarsely crystalline. Light gray, massive. 3% porosity as vugs of unknown origin. Brecciated throughout.

Dolomite; medium crystalline. Light gray, thin to medium-bedded. 10% intercrystalline porosity and fracture porosity. Brecciated at the top.

Dolomite; medium crystalline. Medium gray, massive.

Dolomite; very finely crystalline. Medium gray, massive. 3% vuggy porosity.

Section 2 - Centennial Range



Amsden Formation.

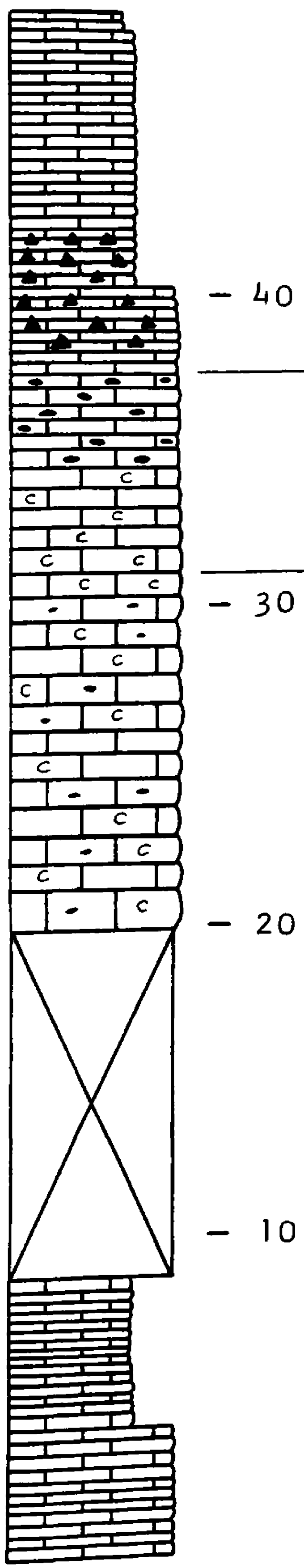
Covered. Exact stratigraphic distance unknown.

Unconformity.

Dolomite; finely crystalline. Light gray, massive. Trace of porosity as small vugs of unknown origin. Cherty at the bottom with a karst breccia at the top.

Dolomite; medium to coarsely crystalline. Light gray, massive. 3% porosity as molds and vugs of unknown origin. Brecciated near the bottom.

Section 2 - Centennial Range



Limestone; brachiopod biosparite (packstone). Medium gray, thin-bedded. Also contains crinoids. Algal-laminated throughout; cherty at the top and brecciated at the bottom.

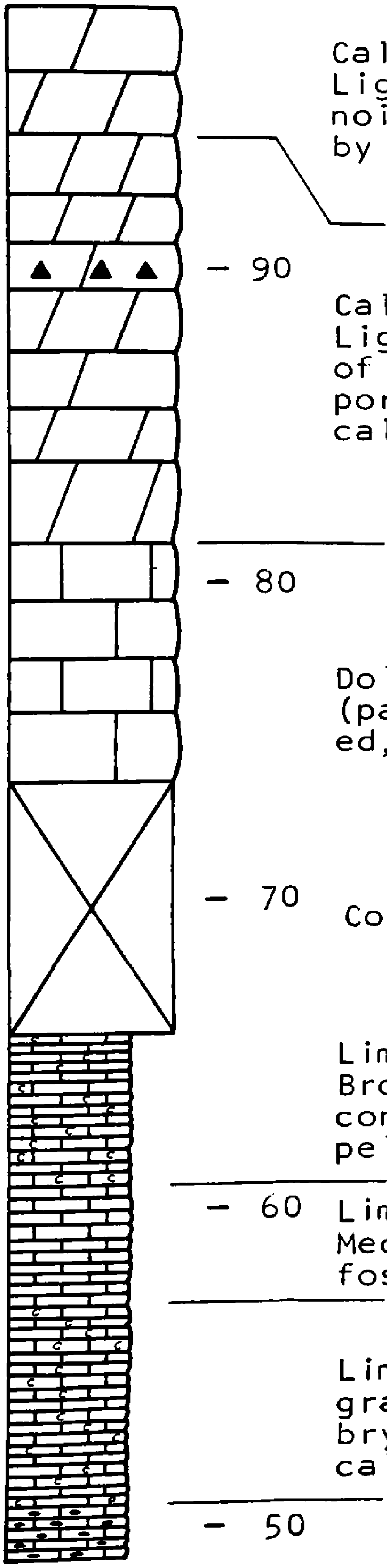
Limestone; biopelmicrite (packstone). Medium gray, medium-bedded. Consists of a peloidal micrite matrix with 25% crinoids, with chert nodules and 2% detrital quartz.

Limestone; crinoidal biomicrite (packstone). Medium gray, medium-bedded. Contains minor amounts of pyrite, chert, and detrital quartz silt. Burrowed and algal-laminated in the middle and grades into mudstone the top 2 m.

Covered.

Lodgepole Formation.

Section 3 - Snowcrest Range



Calcareous dolomite; finely crystalline. Light brown, massive. Contains 15% crinoid particles, dissolved and cemented by sparite. 5% intercrystalline porosity.

Calcareous dolomite, coarsely crystalline. Light brown, massive, with anhedral crystals of dolomite up to 3 mm in size. 20% vuggy porosity, partially filled-in by sparry calcite cement. Brecciated near the top.

Dolomitic limestone; crinoidal biomicrite (packstone). Medium gray, massive. Compacted, with stylolitic porosity.

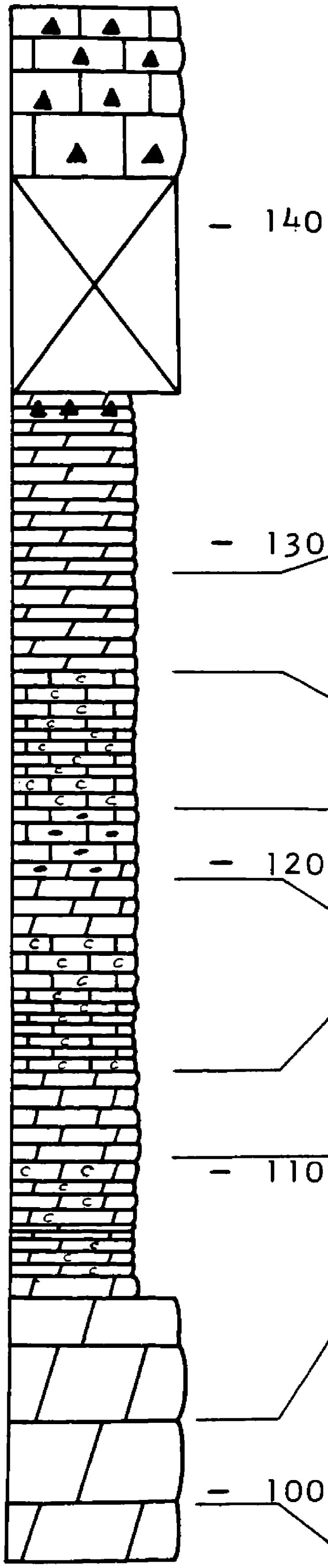
Covered.

Limestone; crinoidal biomicrite (packstone). Brown to reddish-gray, thin-bedded. Also contains brachiopods. Matrix is of .1 mm pellets.

Limestone; fossiliferous micrite (mudstone). Medium gray, thin-bedded. Less than 5% fossil fragments and microspar.

Limestone; biosparite (grainstone). Medium gray, thin-bedded. Consists of crinoids, bryozoans, and pellets cemented by sparry calcite.

Section 3 - Snowcrest Range



Covered.

Dolomite; medium crystalline. Medium brown, thin to medium-bedded. 5% intercrystalline porosity. Becomes more finely crystalline and brecciated toward the top.

Calcareous dolomite; fine to medium crystalline. Light gray, thin to medium-bedded. Contains about 40% crinoids. Slight amount of intercrystalline porosity.

Dolomitic limestone; crinoidal biomicrite (wackestone). Dark gray, thin to medium-bedded.

Calcareous dolomite; finely crystalline. Light brown, thin to medium-bedded. Cherty, with 10% vuggy porosity.

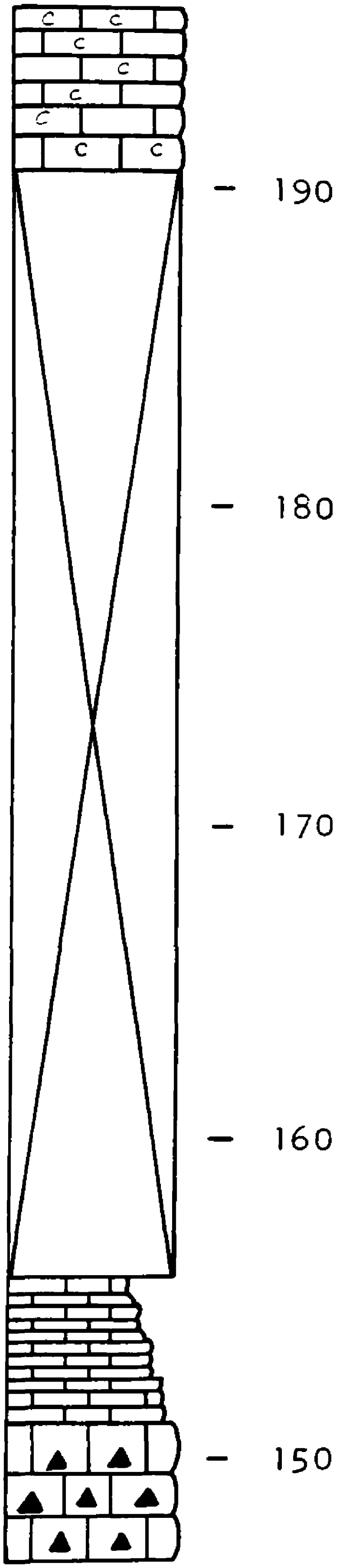
Dolomitic limestone; crinoidal biomicrite (wackestone). Medium gray, medium-bedded.

Calcareous dolomite; finely crystalline. Light gray, thin to medium-bedded. Contains about 40% crinoids.

Dolomite; medium crystalline. Medium brown and massive at the bottom, becoming medium to thin-bedded at the top. Contains 15% dolomitized crinoid particles and has 15% intercrystalline porosity.

Calcareous dolomite; medium crystalline. Light gray, massive. Contains crinoids and dolomitized crinoids, and 5% porosity as vugs and solution-enlarged molds of crinoids. Rock becomes more calcareous and porosity increases toward the top.

Section 3 - Snowcrest Range

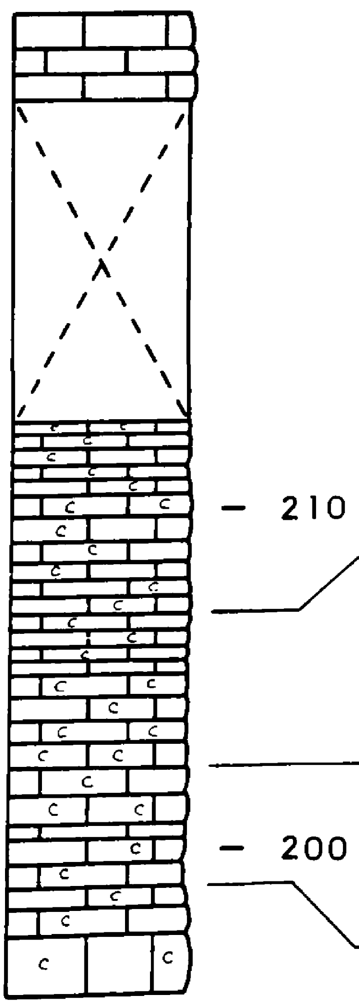


Limestone; crinoidal biomicrite (packstone). Medium gray, medium to thick-bedded. All micrite has been neomorphosed to microspar.

Covered.

Dolomitic limestone; micrite (mudstone). Light to medium gray, thick-bedded and brecciated at the bottom; becomes thin to medium-bedded at the top. Contains 5 to 10% dolomite euhedra.

Section 3 - Snowcrest Range



Big Snowy Group.

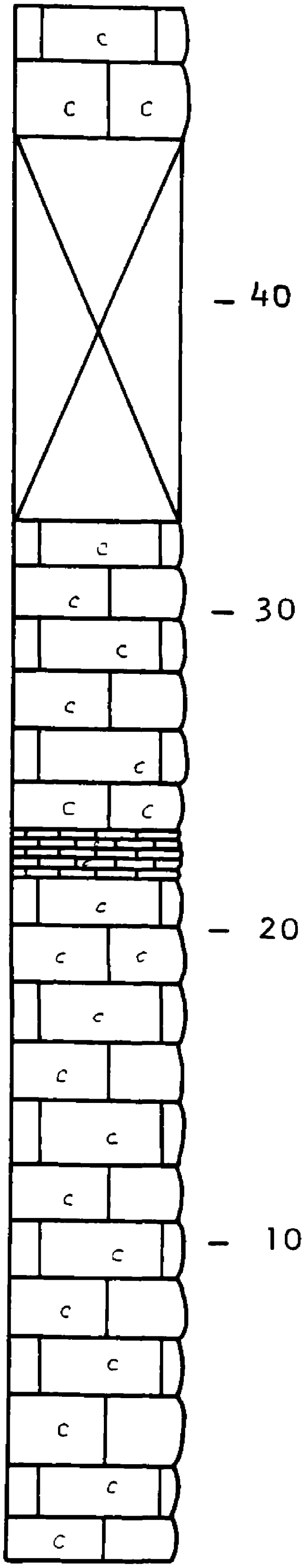
Covered. Exact stratigraphic distance unknown.

Dolomitic limestone; crinoidal pelsparite (grainstone). Dark gray, thin to medium bedded. Compacted texture.

Dolomitic limestone; crinoidal biomicrite (packstone/wackestone). Dark gray, thin to medium bedded. Approximately 30% crinoids in a micrite and microspar matrix.

Dolomitic limestone; crinoidal biosparite (grainstone). Dark gray, thin to medium bedded. Moderately well-compacted, with pressure-solution contacts and broken grains.

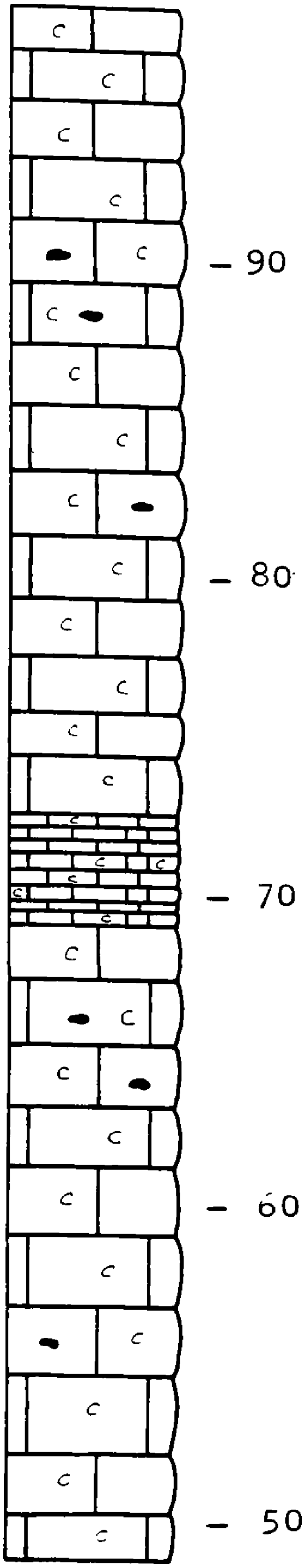
Section 3 - Snowcrest Range



Covered.

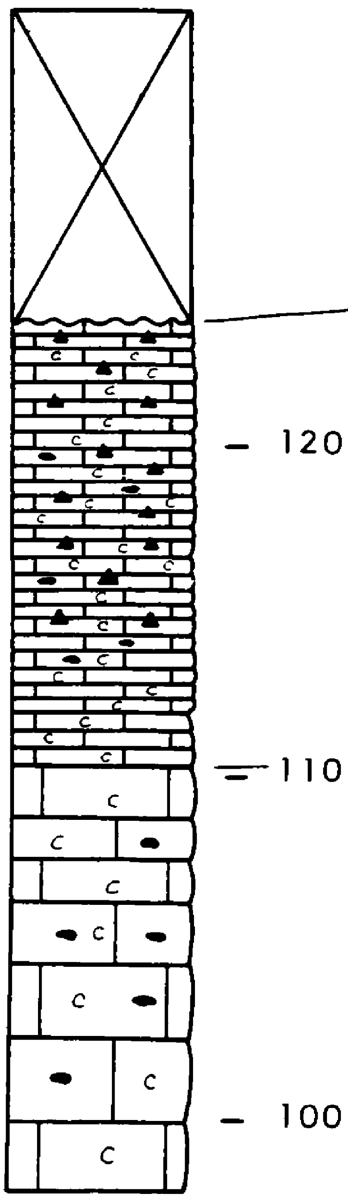
Limestone; crinoidal biomicrite (packstone). Light to medium gray, with occasional thin beds. Consists of 40-50% crinoids in a matrix of micrite, microspar, and pseudospar. Up to 15% porosity the top 5 m as vugs of unknown origin and solution-enlarged fractures. Tectonically deformed throughout.

Section 4 - Camp Creek



Limestone; crinoidal biomicrite (packstone). Light to medium gray, massive, with occasional thin beds. Consists of 40-80% crinoids in a matrix of microspar, with occasional pseudospar and/or micrite. Chert stringers and pods occur occasionally throughout. Tectonically deformed.

Section 4 - Camp Creek



Covered.

Unconformity.

Limestone; crinoidal biomicrite (packstone). Light to medium gray, thin to medium bedded. Consists of 50% crinoids in a matrix of microspar, pseudospar, and micrite. Trace of authigenic quartz replacing some crinoid particles. Brecciated and tectonically deformed throughout.

Section 4 - Camp Creek

Appendix B - Thin-section summary.

Explanation for thin-section summary:

Field Sample -	This number is scratched on the thin-section. It refers to the stratigraphic section (1= Beartooth Mountains, 2=Centennial Range, 3= Snowcrest Range, 4=Camp Creek) and to a particular sampled interval.
U.M.P. #	University of Montana Petrographic Collection number. This number refers to thin sections cataloged in the University of Montana Petrographic collection.
Rock Type	Mdst Lime mudstone Wkst Wackestone Pkst Packstone Gnst Grainstone Dol Dolomite

Field Sample	UMP #	Rock Type	Echinoderms	Brachiopods	Bryozoans	Ostracods	Forams	Ooids	Dolomite	Micrite or microspar	Sparry cement	Authigenic quartz	Detrital quartz	Pyrite	Porosity
1-5		Dol							96						4
1-6		Dol							98		1				1
1-7		Dol							97		2				1
1-8		Dol							99						1
1-9	6975	Dol							99					1	
1-10	6976	Dol							88				12		
1-12		Dol							99						1
1-13	6977	Dol							97				tr	1	2
1-14		Dol	20						64		15				1
1-15	6978	Pkst	30						30	18	20	1	1		
1-16	6979	Wkst	29		1				10	60					
1-17		Dol	20						70	5	5				
1-18		Dol	15						70		15				
1-19		Pkst	40		2				10	28	20				
1-20		Mdst							40	60					tr
1-21	6980	Mdst							20	80					
1-22	6981	Dol							95			tr		tr	5
1-23	6982	Gnst	85		5						8	tr			2
1-24	6983	Pkst	57						40			3			
1-25	6984	Pkst	35	2	2				5	55					
1-26		Wkst	20	1	1				20	45	7	3	3		
1-27		Pkst	40	1	1		1	1	25	5	15		1		
1-28	6985	Pkst	45	1	2		1	1	20	10	20		tr		
1-29	6986	Pkst	35	15					5	40	5				

Field Sample	UMP #	Rock Type	Echino-derms	Brachio-pods	Bryozoans	Ostracods	Forams	Ooids	Dolomite	Micrite or microspar	Sparry cement	Authigenic quartz	Detrital quartz	Pyrite	Porosity
1-30	6987	Gnst	8	1	1			55			35		tr		tr
1-31	6988	Gnst	45		2	2					26	4		1	20
1-32	6989	Mdst							20	77	2		1		
1-33	6990	Dol							95		3				2
1-34	6991	Dol	12	3					82						3
1-35	6992	Dol							90						10
1-36	6993	Pkst	45	3	1	1			5	35	5	5			tr
1-37		Dol							99		1				tr
1-38		Dol							95		5				tr
1-39	6994	Dol							95						5
1-40	6995	Dol							80		20				
1-41	6996	Gnst						80	5		15				
1-42	6997	Gnst	45	5	4	4	1		10		31				tr
1-43	6998	Gnst	40				tr	45			15				
1-43	6998	Mdst	tr	tr						99					
1-44	6999	Gnst	83	2							15				
2-8	7000	Pkst	40							20	30				10
2-9	7001	Gnst	60					15			25				
2-10	7002	Gnst	60					15			25				
2-11	7003	Gnst	40								60				
2-12	7004	Dol							78		20				2
2-13		Dol							82		8				10
2-14		Dol							83		12				5
2-15		Dol							84		4				12
2-16		Dol							97						3
2-17	7005	Dol	10						80		10				

Field Sample	UMP #	Rock Type	Echino-derms	Brachio-pods	Bryozoans	Ostracods	Forams	Ooids	Dolomite	Micrite or microspar	Sparry cement	Authigenic quartz	Detrital quartz	Pyrite	Porosity
2-18	7006	Dol							75		15				10
2-19	7007	Dol							90		5				5
2-20	7008	Dol							83		15				2
2-21		Dol							84		10				6
2-22	7009	Dol							80						20
2-23		Dol							80		10				10
2-24	7010	Dol							80		13				7
2-25		Dol							88		10				2
2-26	7011	Dol							75		5				20
2-27	7012	Dol							82		15				3
2-28	7013	Dol							96						4
2-29		Dol							87		5				8
2-31		Dol							82		3				15
2-32	7014	Dol							90						10
2-33		Dol							98						2
2-34	7015	Dol							90						10
2-35		Dol							100						
2-36		Chert										100			
2-37	7016	Dol							80						20
2-38		Dol							98						2
2-39		Dol							96		3				1
2-40		Dol							100						
2-41		Dol							97						3
2-42		Dol							96						4

Field Sample	UMP #	Rock Type	Echino-derms	Brachio-pods	Bryozoans	Ostracods	Forams	Ooids	Dolomite	Micrite or microspar	Sparry cement	Authigenic quartz	Detrital quartz	Pyrite	Porosity
2-43		Dol							80						20
2-44	7017	Dol							97						3
2-45		Dol							97						3
2-46	7018	Dol							60		25				15
2-47	7019	Dol							83						17
2-48	7020	Dol							97						3
2-49	7021	Dol							95						5
2-50		Dol							100						
2-51		Dol							75		5				20
2-52	7022	Dol							100						tr
2-53	7023	Dol							57		tr	40			3
2-54	7024	Dol							93						7
2-55	7025	Dol							100						
2-56		Dol							100						
2-57		Dol							93		5				2
2-58		Dol							75						25
2-59		Dol							96						4
2-60		Dol							97		tr				3
2-61		Dol							100						tr
3-9	7026	Pkst	50	8	2	1				27	10	1	1	tr	
3-10	7027	Pkst	45	3	2	1				30	2	1	3	tr	
3-11	7028	Pkst	40	15	5	3				27	7	3		tr	
3-12	7029	Wkst	26	2						70			2		
3-13	7030	Mdst	2	1	1	1				80	5	10			
3-14	7031	Pkst	6	25		4				15	50				

Field Sample	UMP #	Rock Type	Echino-derms	Brachio-pods	Bryozoans	Ostracods	Forams	Ooids	Dolomite	Micrite or microspar	Sparry cement	Authigenic quartz	Detrital quartz	Pyrite	Porosity
3-15	7032	Pkst	5	30	5	2				20	38				
3-16	7033	Pkst	35	15						50					
3-17	7034	Gnst	45	15	2	1	1				35	1		tr	
3-18		Mdst								100					
3-19	7035	Pkst	50	10	1	1	1			33		1	3		
3-20	7036	Gnst	55	5					5		35				tr
3-21		Dol							85		5				10
3-22		Dol							90						10
3-23	7037	Dol							70		10				20
3-24	7038	Dol							75		5				20
3-25		Dol							85		10				5
3-26	7039	Dol							85		10				5
3-27	7040	Dol							85		5				10
3-28	7041	Dol							85						15
3-29	7042	Dol	40			tr			50		5	tr			
3-30		Pkst	35						15	30	20				
3-31		Pkst	40						15	30	15				
3-32	7043	Dol							73		15	tr		2	10
3-33		Pkst	65						20	15					
3-34	7044	Pkst	45						50		1		2		2
3-35	7045	Dol							90		5	tr			5
3-36	7046	Dol							80		5				15
3-38	7047	Mdst							8	92					
3-39	7048	Mdst							7	75	10				8

Field Sample	UMP #	Rock Type	Echino-derms	Brachio-pods	Bryozoans	Ostracods	Forams	Ooids	Dolomite	Micrite or microspar	Sparry cement	Authigenic quartz	Detrital quartz	Pyrite	Porosity
3-40	7049	Pkst	40						5	50					5
3-41	7050	Gnst	65	8	1	1	1		10		14				
3-42		Wkst	30						10	60					
3-43	7051	Gnst	57	4	2	1	1	tr	10	tr	15				
4-1	7052	Pkst	55							45					
4-2	7053	Pkst	50							48		2			
4-4	7054	Pkst	60							40					
4-11	7055	Pkst	60							40					
4-12		Pkst	50							50					
4-13	7056	Pkst	80							10		10			
4-14		Pkst	65							35		tr			
4-15		Pkst	50							50					
4-16		Pkst	55							65					
4-17	7057	Pkst	50							50					

Appendix C - Shallowing-upward cycles.

Explanation for appendix C:

Mdst	Mudstone
Wkst	Wackestone
Pkst	Packstone
Gnst	Grainstone
Dol	Dolomite
Cov	Covered

Symbols for fossil particles

Echinoderms

Bryozoans

Brachiopods

Ostracods

Calcareous algae

Forams

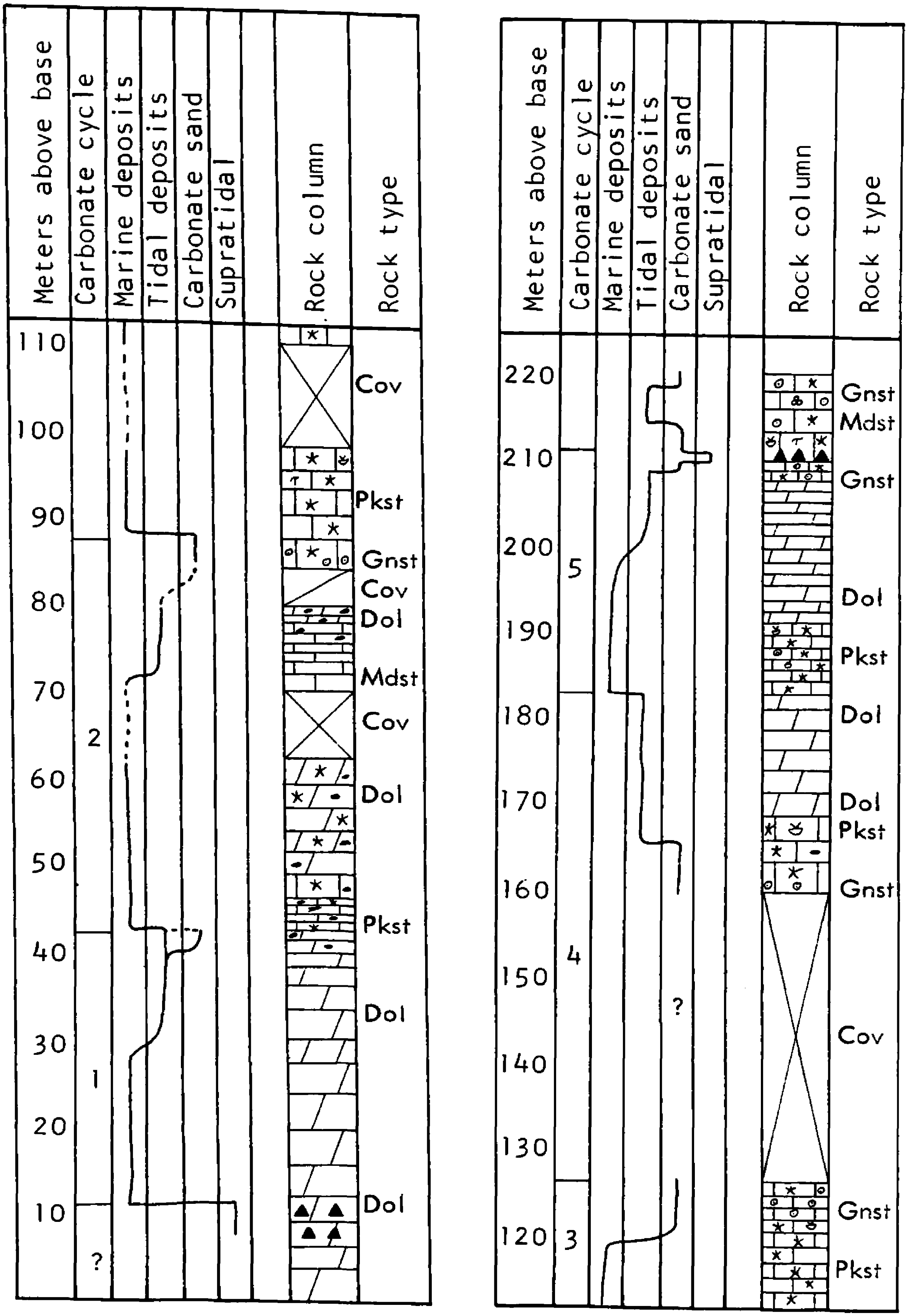
Gastropods

Ooids

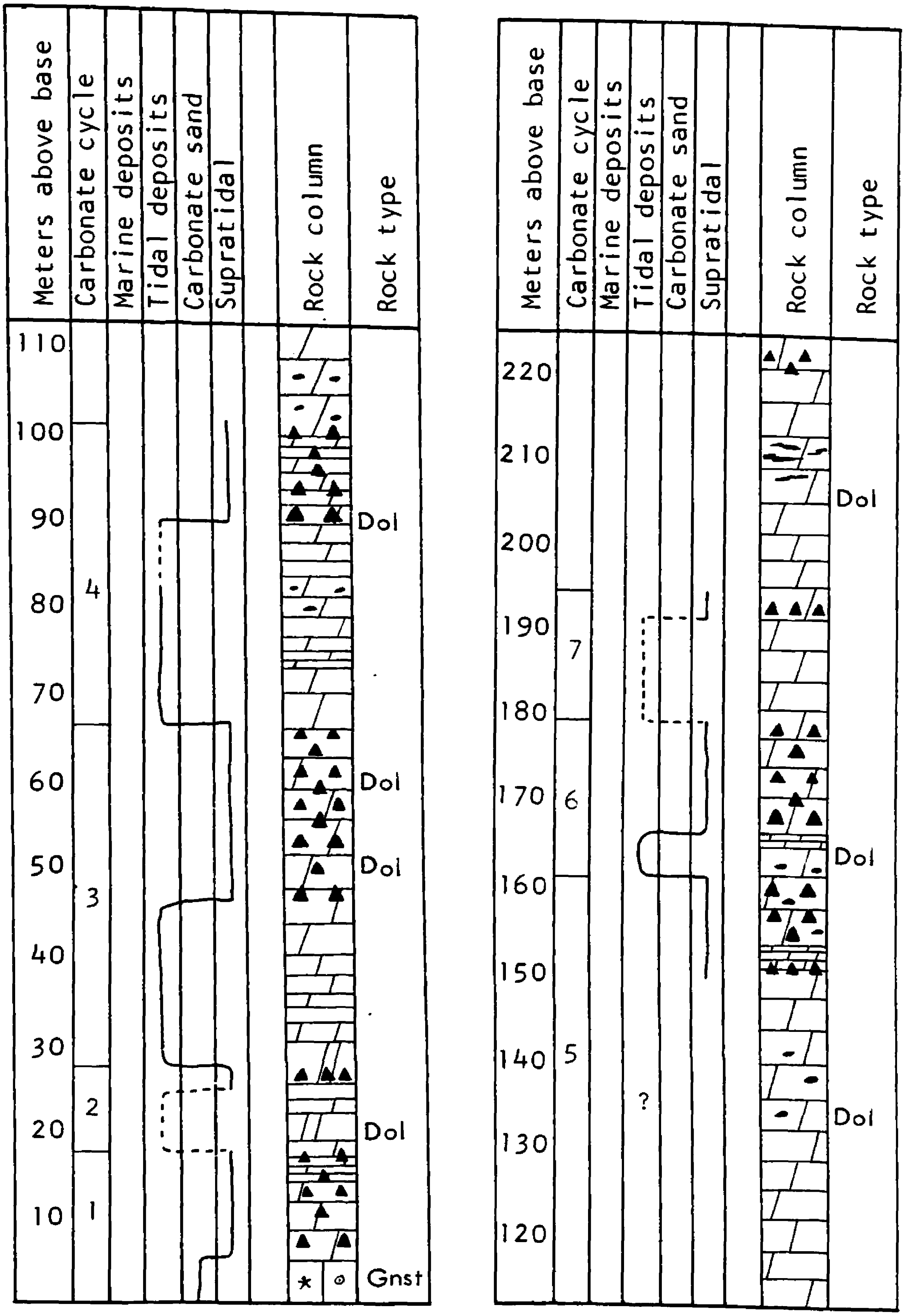
Solitary corals

Chert

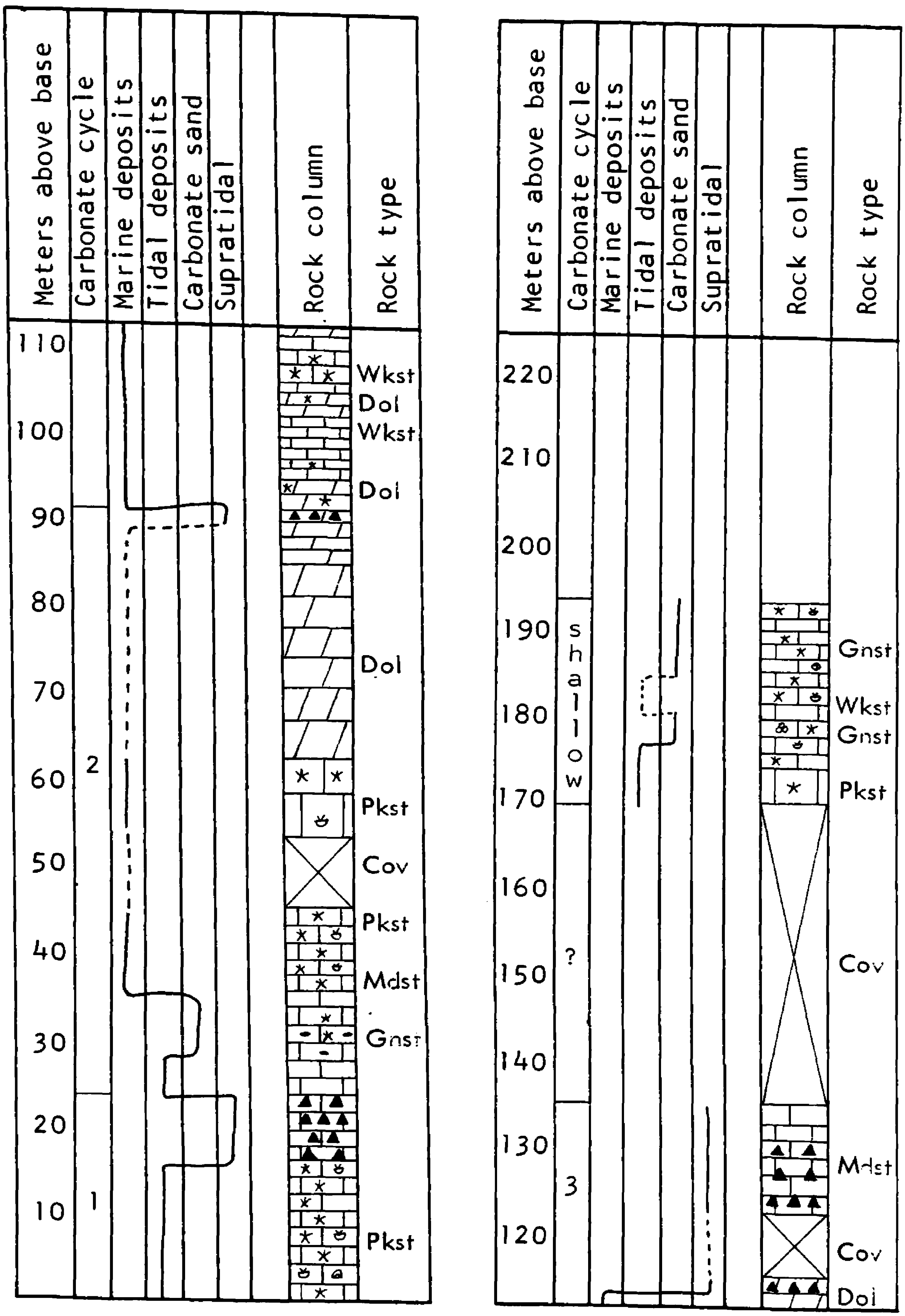
Breccia



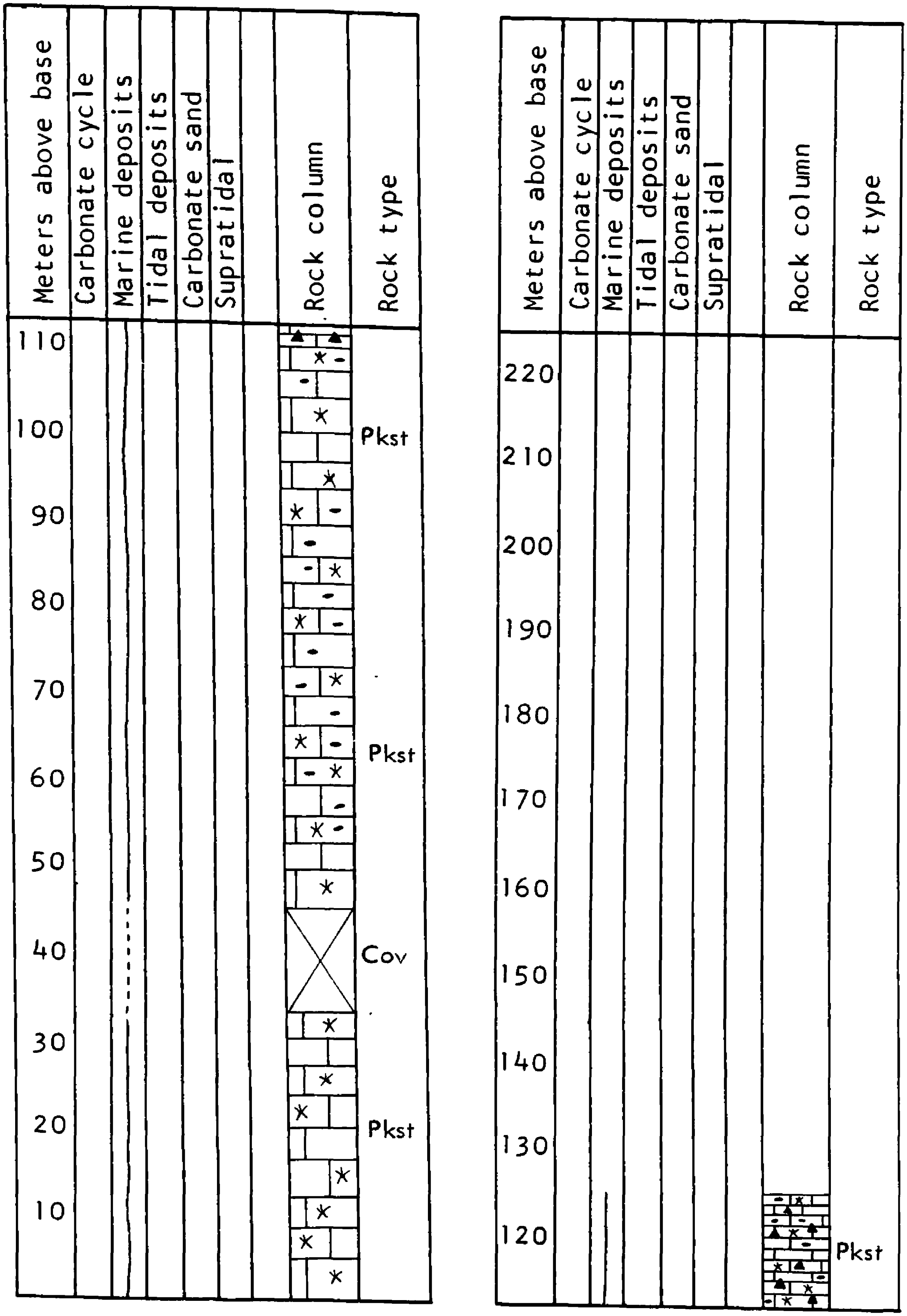
Appendix C. Shallowing-upward cycles in the Mission Canyon Formation. Beartooth Mountains section.



Appendix C. Shallowing-upward cycles (continued). Centennial Range section.



Appendix C. Shallowing-upward cycles (continued). Snowcrest Range section.



Appendix C. Shallowing-upward cycles (continued). Camp Creek section.